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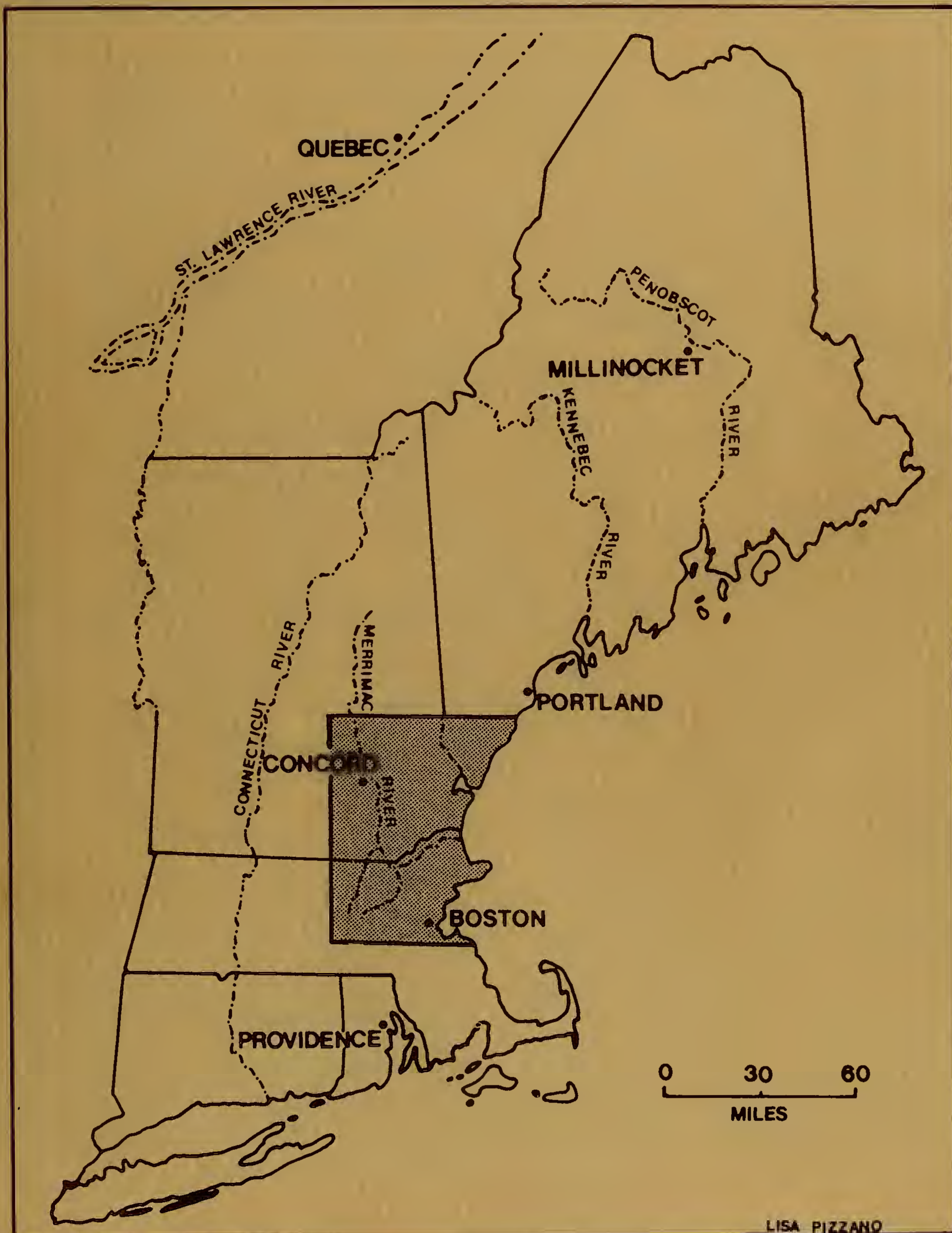
Regional Geology and Tectonic History of Southeastern New England; Quarternary Geology and Geomorphology

Barosh, Patrick J.

Hanson, Lindley S.

Geology of the Coastal Lowlands Boston, MA to Kennebunk, ME

76th ANNUAL NEW ENGLAND INTERCOLLEGIATE GEOLOGIC CONFERENCE, 1984



Edited by Lindley S. Hanson

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COASTAL LOWLANDS FROM BOSTON TO KENNEBUNK AND INLAND
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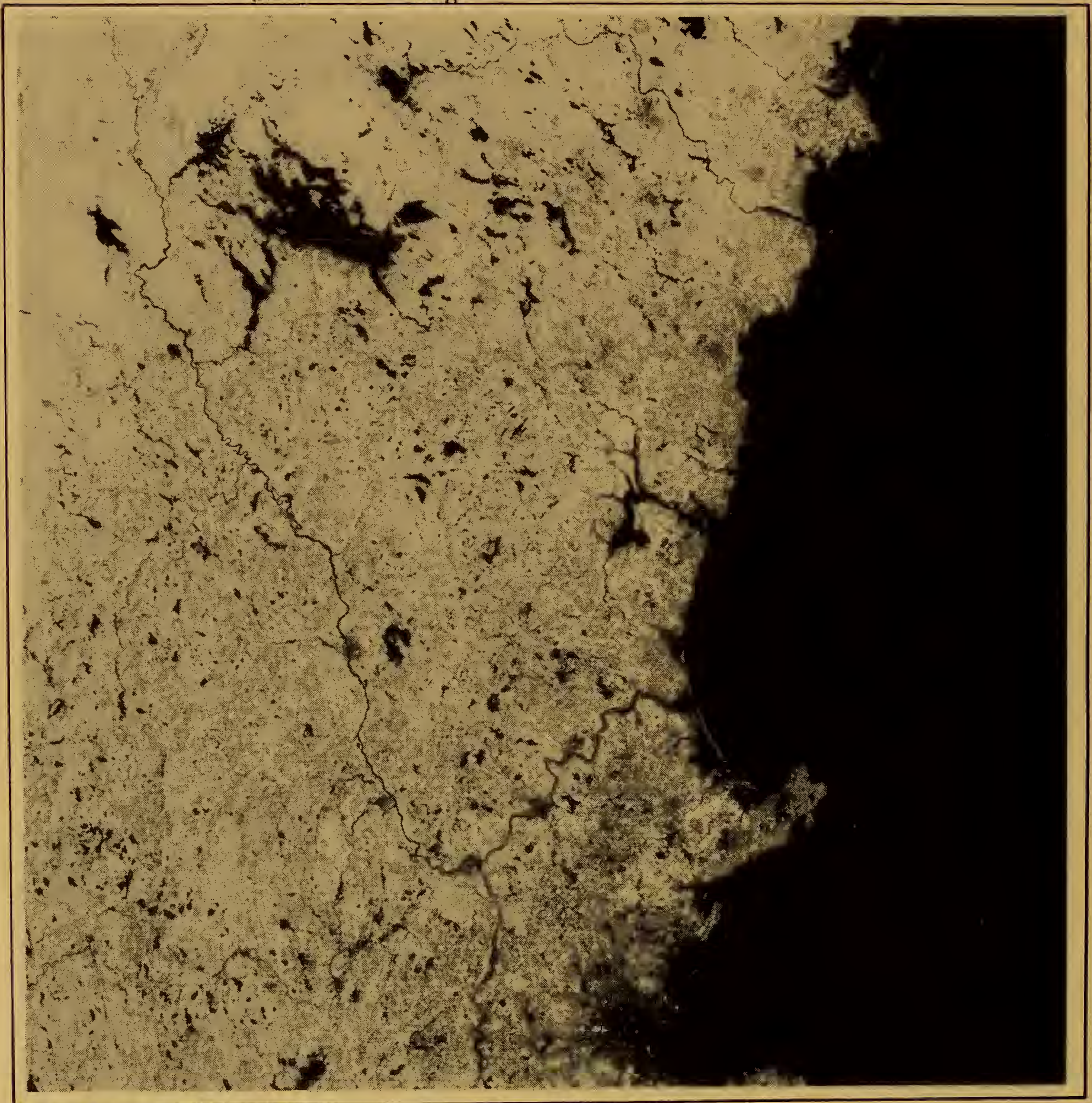

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GEOLOGY OF THE COASTAL LOWLANDS, BOSTON TO KENNEBUNK, MAINE

Edited by

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Salem, Massachusetts 01970

76th ANNUAL MEETING
NEW ENGLAND INTERCOLLEGIATE GEOLOGICAL CONFERENCE

Danvers, Massachusetts

October 12, 13, 14, 1984

Host

Salem State College
Salem, Massachusetts 01970

76TH ANNUAL MEETING
NEW ENGLAND INTERCOLLEGIATE GEOLOGICAL CONFERENCE

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PREFACE

THE NEW ENGLAND INTERCOLLEGIATE GEOLOGICAL CONFERENCE

The New England Intercollegiate Geological Conference (NEIGC) was organized in the fall of 1901 by William Morris Davis, who led a trip to the Westfield River area in Massachusetts (see Table 1 and Figure 1). Although no record of the first conference exists, it appears that there was a single field trip, perhaps lasting more than one day, and with Davis as the leader. Legend also has it that only a few people attended. With the 1984 meeting, the NEIGC will not have met during 8 years; 6 years during World Wars I and II, and an unexplained 2-year hiatus between 1912 and 1915. The present Secretary understands such a break, however, given the difficulty in finding people to take on the conference organization.

By 1927 or 1928, at least 2 separate field trips (bedrock and surficial geology) were being offered. With only a few exceptions, multiple field have been the rule. So many field trips were being offered by 1959 that only the conference organizers of that and subsequest meetings can be listed in Table 1; the number of trip leaders now exceed the number of people attending the early meetings.

The sole purpose of NEIGC has always been to conduct field trips in areas of recent geologic investigation. The 1984 meeting in northeastern Massachusetts and southwestern Maine certainly is no exception to this rule, with a new map of the bedrock of Massachusetts published in 1983, E-an Zen editor, and a new map of the surficial geology of the Commonwealth in the works, Byron Stone, editor. The State of Maine has two new maps in press; a map of the bedrock edited by Boone, Hussey and Osberg, and a surfical geology map edited by Borns and Thompson.

The early trip leaders and organizers of NEIGC were academics, but for a long time geologists from the United States Geological Survey, the Geological Survey of Canada, from numerous state geological surveys and from industry have all assisted in conducting field trips and organizing the conferences. The field trip leaders of the early conferences presented the geology and road logs in hand outs. These have been replaced by guidebooks that are now important sources of information for the geology of the northeast, and are shelved in some of the major libraries of the country.

The NEIGC has met outside of New England 9 times, indicating that the geology of New England extends far beyond its borders. Five meetings have been in New York, 3 meetings were held in the Province of Quebec, and 1 meeting was held in New Brunswick. In addition, many other conferences held near its borders have had individual field trips outside of New England. With the 1984 meeting in Danvers, the NEIGC has met 26 times in Massachusetts, 11 times each in Maine and Connecticut, 8 times in New Hampshire, 6 times in Rhode Island, and 5 meetings have been held in Vermont.

Future meetings of NEIGC are planned as follows: 1985 in New Haven, Connecticut, Robert Tracy, organizer; 1986 in Lewiston, Maine, Don Newberg and Marc Ratelle, organizers; and the 79th annual meeting in 1987 in Norwich, Vermont, with Fred Larsen and Dave Westerman as the organizers.

D. W. Caldwell, Secretary
NEIGC

LOCATIONS OF PAST NEIGC MEETINGS
QUEBEC
77
FREDRICTON
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21
15
39,53
07,24
32
15
58
82
47
63
03,06,12,20
81
32
17
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50
Miles
08
37
numbers refer to conference dates in this century
DUANE MATTERS

TABLE 1

LIST OF MEETINGS
OF THE
NEW ENGLAND INTERCOLLEGIATE GEOLOGICAL CONFERENCE

Meeting	Year	Location	Organizer
1st	1901	Westfield River Terrace, Mass.	Davis
2nd	1902	Mount Tom. Massachusetts	Emerson
3rd	1903	West Peak, Meriden, Conn.	Rice
4th	1904	Worcester, Massachusetts	Emerson
5th	1905	Boston Harbor and Nantasket	Johnson, Crosby
6th	1906	Meriden to East Berlin, Conn.	Gregory
7th	1907	Providence, Rhode Island	Brown
8th	1908	Long Island, New York	Barrell
9th	1909	North Berkshires, Mass.	Cleland
10th	1910	Hanover, New Hampshire	Goldthwait
11th	1911	Nahant and Medford, Mass.	Lane, Johnson
12th	1912	Higby-Lamentation Blocks	Rice
13th	1915	Waterbury to Winsted, Conn.	Barrell
14th	1916	Blue Hills, Massachusetts	Crosby, Warren
15th	1917	Gay Head and Martha's Vienhead	Woodworth, Wigglesworth
16th	1920	Lamentation and Hanging Wall	Rice, Foye
17th	1921	Attleboro, Massachusetts	Woodworth
18th	1922	Amherst, Massachussetts	Antevs
19th	1923	Beverly, Massachusetts	Lane
20th	1924	Providence, Rhode Island	Brown
21st	1925	Waterville, Maine	Perkins
22nd	1926	New Haven, Connecticut	Longwell
23rd	1927	Worcester, Massachusetts	Perry, Little, Gordon
24th	1928	Cambridge, Massachusetts	Billings, Bryan, Mather
25th	1929	Littleton, New Hampshire	Crosby
26th	1930	Amherst, Massachusetts	Loomis, Gordon
27th	1931	Montreal, Quebec	O'Neill, Clark, Gill
28th	1932	Providence, Rhode Island	Brown
29th	1933	Williamston, Massachusetts	Cleland, Perry, Knopf
30th	1934	Lewiston, Maine	Fisher, Perkins
31st	1935	Boston, Massachusetts	Morris, Pearshall, Whitehead
32nd	1936	Littleton, New Hampshire	Billings, Hadley, Cleaves
33rd	1937	New York City & Duchess Co.	O'Connell, Kaye, Fluhr, Balk
34th	1938	Rutland, Vermont	Bain
35th	1939	Hartford & Conn. Valley	Troxell, Flint, Longwell
36th	1940	Hanover, New Hampshire	Goldthwait, Denny, Stoiber
37th	1941	Northampton, Massachusetts	Balk, Jahns, Lochman
38th	1946	Mt. Washington, N.H.	Billings
39th	1947	Providence, Rhode Island	Quinn
40th	1948	Burlington, Vermont	Doll
41st	1949	Boston, Massachusetts	Nichols, Billings, Schrock
42nd	1950	Bangor, Maine	Trefethen, Raisz
43rd	1951	Worcester, Massachusetts	Lougee, Little
44th	1952	Williamston, Massachusetts	Perry, Foote, McFadyen
45th	1953	Hartford, Connecticut	Flint, Gates, Peoples
46th	1954	Hanover, New Hampshire	Elston, Washburn, Lyons
47th	1955	Tigonderoga, New York	Rodgers, Walton, Bartolome
48th	1956	Portsmouth, New Hampshire	Novotny, Billings, Chapman
49th	1957	Amherst, Massachusetts	Bain, Johannson, Rice

Meeting	Year	Location	Organizers
50th	1958	Middleton, Connecticut	Rosenfield, Eaton, Sanders
51st	1959	Rutland, Vermont	Zen
52nd	1960	Rumford, Maine	Griscom, Milton, Caldwell
53rd	1961	Montpelier, Vermont	Doll
54th	1962	Montreal, Quebec	Clark
55th	1963	Providence, Rhode Island	Quinn
56th	1964	Chestnut Hill, Massachusetts	Skehan
57th	1965	Brunswick, Maine	Hussey
58th	1966	Katahdin, Maine	Caldwell
59th	1967	Amherst, Massachusetts	Robinson, Drake, Foose
60th	1968	New Haven, Connecticut	Orville
61st	1969	Albany, New York	Bird
62nd	1970	Rangeley Lakes, Maine	Boone
63rd	1971	Concord, New Hampshire	Lyons, Stewart
64th	1972	Burlington, Vermont	Doolan, Stanley
65th	1973	Fredericton, New Brunswick	Greiner
66th	1974	Orono, Maine	Osberg
67th	1975	Great Barrington, Mass	Ratcliffe
68th	1976	Boston, Massachusetts	Cameron
69th	1977	Quebec City, Quebec	Beland, LaSalle
70th	1978	Calais, Maine	Ludman
71st	1979	Troy, New York	Friedman
72nd	1980	Presque Isle, Maine	Roy, Naylor
73rd	1981	Kingston, Rhode Island	Boothroyd, Hermes
74th	1982	Storrs, Connecticut	Joesten, Chestnut, Quarrier
75th	1983	Greenville and Millinocket, Maine	Caldwell and Hanson
76th	1984	Danvers, Massachusetts	Hanson

ACKNOWLEDGEMENTS

I would like to thank the Administration of Salem State College, for their continued support and assistance in the organization of this conference, and the Salem State Department of Geography, for their assistance in the production of illustrations. I would also like to extend my gratitude to all field trip leaders for their participation and efforts to make this a memorable conference.

Lindley Hanson

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QUATERNARY GEOLOGY AND GEOMORPHOLOGY

REGIONAL GEOLOGY AND TECTONIC HISTORY OF SOUTHEASTERN NEW ENGLAND

by

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Boston College

Introduction

Southeastern New England contains some of the most interesting, varied, and complex geology of all of North America. It lies astride the eastern edge of the Paleozoic Appalachian orogenic belt and offers a glimpse of this margin and the structure to the east of this belt that may not be seen elsewhere on the Atlantic coast of the United States. The boundary zone along the margin is the greatest structural zone known in New England and apparently represents a zone of collision between a Paleo-North American and Paleo-African plates. Structure and deposits related to the rifting of the present North Atlantic basin are also displayed in the region, as well as a wide variety of features and materials resulting from Pleistocene glaciation. Earthquake activity indicates it is still a tectonically active region. Numerous kinds of structures are present in the rock, which represents all periods from Precambrian to Quarternary with only the possible exception of the Mississippian.

Repeated glaciation during the Pleistocene stripped off the overbunden from the bedrock and then haphazardly recovered most of it with a wide variety of deposits. Since the retreat of the ice, many hollows left in the topography have filled with soft sediments and peat and the rising sea has formed coastal marshes and beaches. The rising sea level, regional tilt to the south due to post-glacial rebound and local tectonic subsidence have combined to create a very interesting and complex history along the coast.

The extreme variety in rock and surficial deposits and their structure make the region one of the most challenging anywhere for all types of geologists, as nearly every site is different. The growing need to understand more about water movement, both for supply and hazardous waste sites and ground conditions for an ever-expanding variety of construction requires more detailed knowledge of the geology than ever before.

A virtual explosion of new information has become available for the region over the past 15 years that has radically changed the earlier concepts of the region. Most of the data is still unpublished. It is derived mainly from the mapping of the Boston Office of the U. S. Geological Survey under L. R. Page and M. H. Pease, Jr., data from the New England Seismotectonic Study, sponsored by the U. S. Nuclear Regulatory Commission, and the recent work of the office of Marine Geology of the U. S. Geological Survey. This detailed mapping employed modern statigraphic studies, geophysical surveys, and radiometric age dating from well controlled samples, to great advantage. The understanding of the geology of the region is thus, undergoing many changes resulting from these new data and their ramifications. Much of the recent geologic literature on the area, however, reflects this understanding unevenly and is often contradictory. The field trips presented in this volume may not be entirely consistent with one another, but they maintain the valuable tradition of the New England Intercollegiate Geologic Conference of providing an opportunity of seeing the geology in the field. The participants may or may not agree with the interpretation presented, but they will be rewarded with some stimulating discussion and a greater knowledge of the geology of the region.

This report will briefly describe the geology of the region and its tectonic history to provide a background for the field trips. The following description may vary considerable from state geologic maps of Massachusetts (Zen, 1983), Connecticut (Rodgers, 1982), and New Hampshire (Billings, 1955), as considerable more recent work has been drawn upon for this report. This summary draws upon the mapping of a great many people whose work could not be all cited in such a short paper; most of this is listed in Barosh and others (1977). The references cited tend to represent more summary reports and newer findings. Much of the recent work contained herein was supported by the U. S. Nuclear Regulatory Commission under Contract Number AT (49-24) - 0291.

Structural Framework

Southeastern New England is formed of two vastly different geologic terrains separated by a wide zone of steeply west-dipping thrust faults, the Nashoba Thrust Belt (Barosh and others, 1977) (Fig.1). No rocks have been correlated across this belt, nor does the meta-volcaniclastic rock within it occur on either side. The Nashoba Thrust Belt forms the largest structural discontinuity known in the northeastern United States and apparently separates blocks belonging to Paleo-African and North American plates that collided here. This belt passes just to the northwest of Boston. Southeast of the thrust belt is a largely Precambrian granite terrain, the Southeast New England Platform, into which the Boston and other basins have been dropped. To the northwest lies the Sturbridge Geocline of pre-Ordovician meta-sedimentary rock with a block of Late Silurian (?) - Early Devonian (?) rock, caught between thrust fault zones. Remnants of Pennsylvanian and Late Triassic - Early Jurassic rock in grabens and fault slivers are present mainly in the Southeastern New England Platforms. All the terrains are overlapped just off-shore by Cretaceus and Tertiary deposits that form the submerged northward extension of the Atlantic Coastal Plain.

Southeast New England Platform

The Southeast New England Platform (Fig. 1) consists of a late Precambrian batholithic complex and associated metasedimentary and metavolcanic rocks that were intruded by plutons and covered by sediments at various times during the Paleozoic. The covering sediments are preserved in basins that are largely fault bounded (Fig. 1). These include the Boston Basin, which contains latest Precambrian to Middle Cambrian conglomerate, argillite, and volcanic rock (Kaye and Zartman, 1980; Kaye, 1981) and perhaps Orderician volcanic rock, the Narragansett and Norfolk Basins which contain Pennsylvanian stream deposits that locally overlie trilobite-bearing Cambrian limestone and phyllite (Shaler and

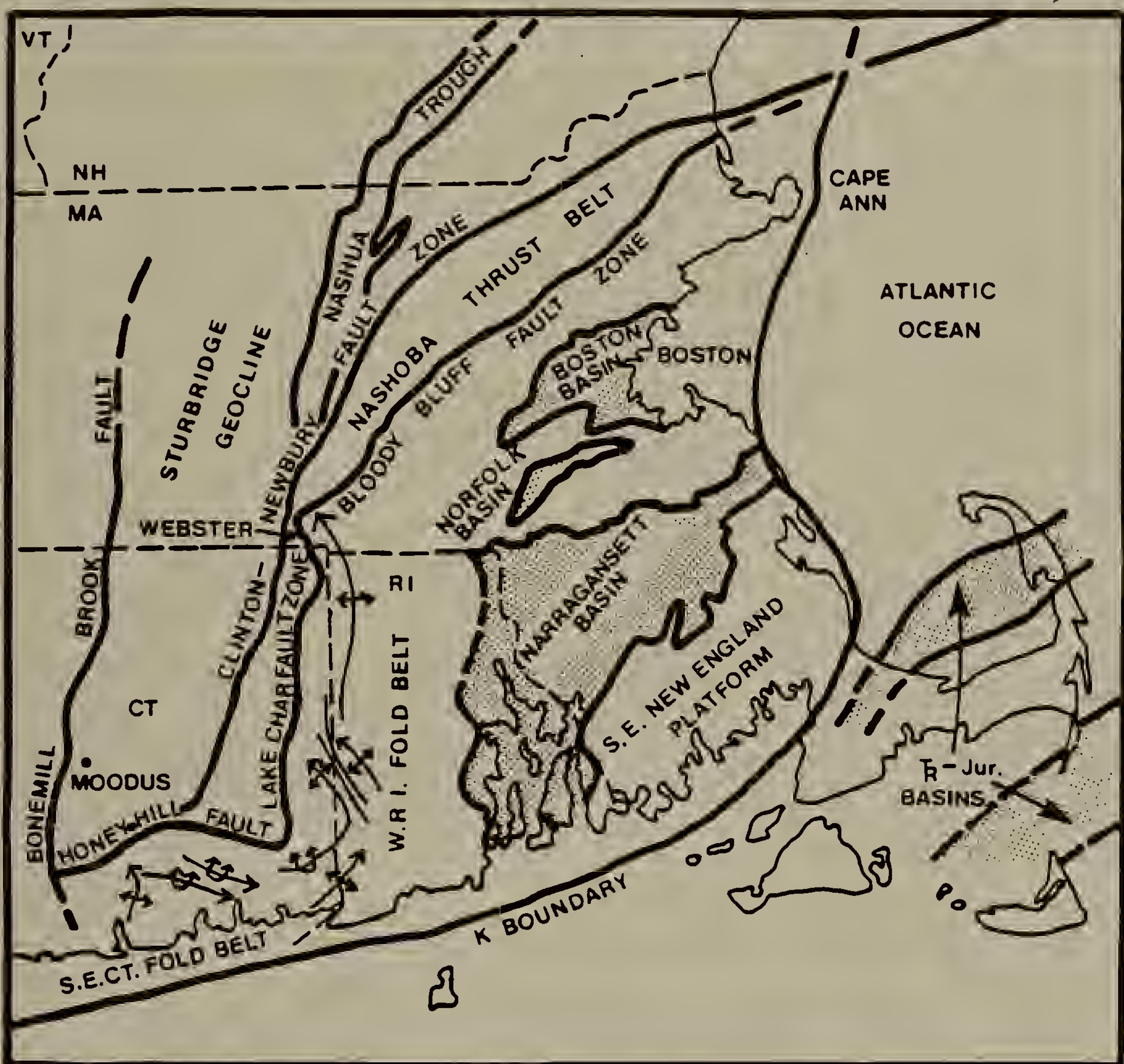

Fig. 1 Map of southeastern New England showing major tectonic provinces and structures.

others, 1899); and many off-shore and rare on-shore basins of Late Triassic to Early Jurassic sandstone, siltstone, and basalt (Ballard and Uchupi, 1975). Locally common are Paleozoic and Mesozoic dikes of diabase and lamprophyre.

The batholithic complex contains xenoliths and large pendants of metasedimentary and metavolcanic rock, that appear to be more abundant along its western side. These rocks form a sequence of quartzite and interbedded schist grading up into amphibolite and some metabasalt. They are broadly referred to as the Westboro Formation, Plainfield Formation or Blackstone Series in Massachusetts, Connecticut and Rhode Island, respectively (the Westboro Formation is examined northwest of Boston on trip B3 by Bailey). West and northwest of Boston, the quartzite is succeeded by a sequence of fine-grained, thin-bedded tuffaceous metasediments, including some very well laminated tuffs, that form the Middlesex Fells Volcanic Complex, the Greenleaf Mountain Formation, Burlington Formation and some unnamed gneiss and quartzite (Bell and Alvord, 1976).

The composition of the late Precambrian complex ranges from quartz-rich alaskite to diorite or gabbro (Hermes, Gromet and Zartman, 1981). Light colored granodiorite and quartz monzonite are common. The Dedham granodiorite is one of the more widespread plutons and at least several others are present. The division between plutons and local variations within them is not everywhere known. Most are nonfoliated and many are remarkably fresh looking, although the Dedham has undergone alteration in most places and the original light gray rock consequently has developed a pinkish cast.

Rocks of the batholithic complex are in the range of 600 to 620 m.y. in age (Zartman and Naylor, in press; Galloway 1973; Smith 1978, Hermes, Gromet and Zartman, 1981), and have not been definitely found elsewhere in New England (the Massabesic Gneiss problem is discussed below). However, this complex is thought to correlate with late Precambrian plutons in the Avalon Peninsula in Newfoundland and together form a distinctive structural belt referred to as the Avalon Zone (Rast and others, 1976) (part of the batholithic complex north of Boston is visited on trip C3 by Smith and Hon).

The complex is more deformed towards the edge of the Nashoba Thrust Belt, marked by the Lake Char and Bloody Bluff fault systems, and may be strongly foliated, sheared, and folded adjacent to it (Figs. 1 and 2). The rock in Rhode Island, west of the Narragansett Basin, forms a complex broad north-trending dome, the Rhode Island Dome (Fig. 2). The general domal structure is defined by the attitude of bedding in the intruded Precambrian metamorphic rock. These strata lie mainly at the edges of the intrusive-cored dome; they dip to the west in eastern Connecticut, to the north in adjacent Massachusetts, and to the northeast in northeast Rhode Island.

This general domal structure is cut by numerous faults and is distorted by smaller folds. A series of north-trending, north-plunging folds, designated the West Rhode Island Fold Belt (Barosh 1976; Barosh and Hermes 1981; and Hermes, Barosh and Smith, 1981), lies along the Connecticut-Rhode Island border (Figs. 2 and 3). These folds are broad

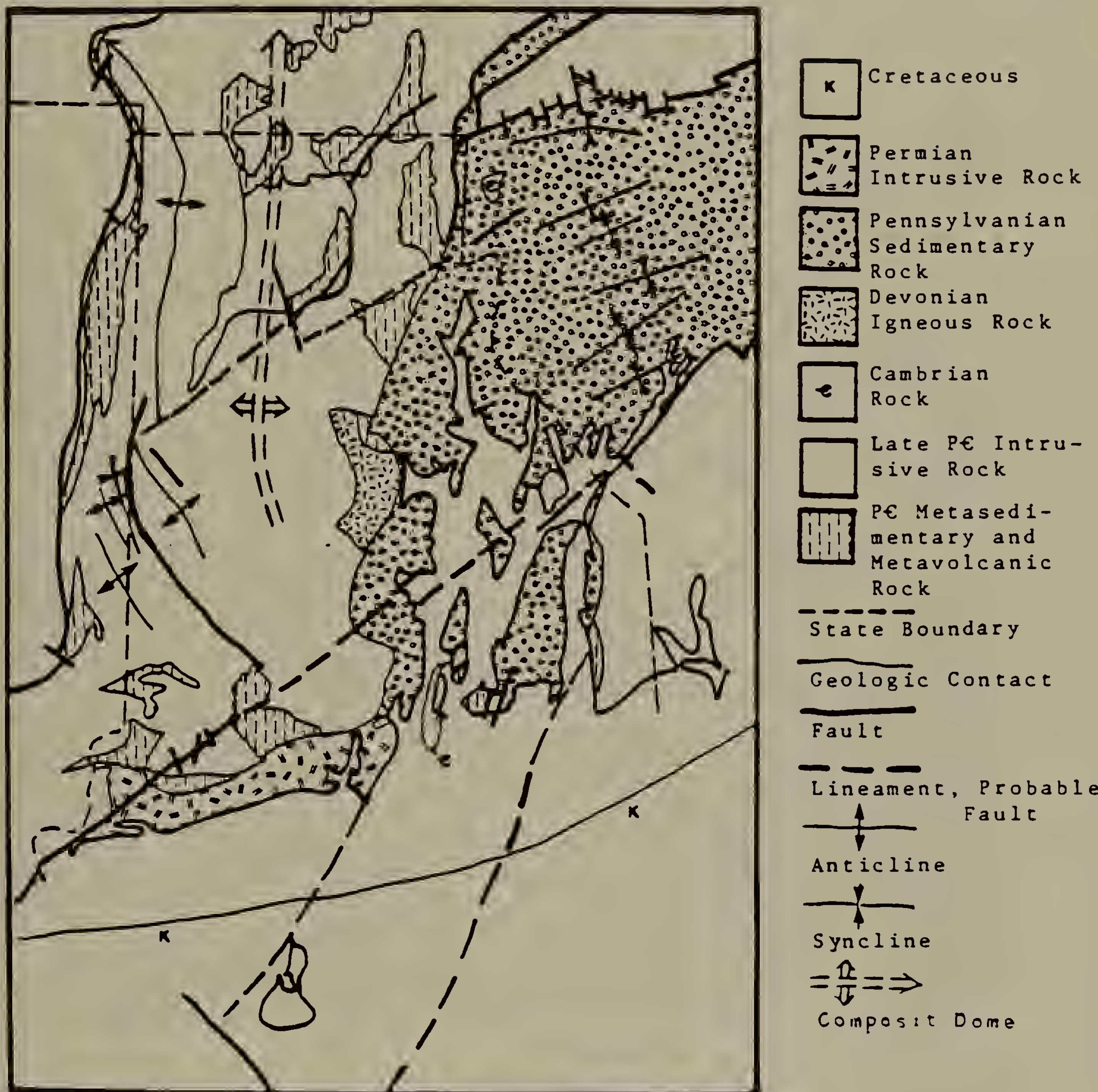

Fig. 2 Sketch map of Rhode Island and vicinity showing major structural features of the southeast New England Platform, (Barosh and Hermes, 1981)

and open in the north, but become progressively more compressed to the south where they are overturned and broken by thrust faults. The faults tend to cut out the synclines. The western part of the fold belt swings southwest and west approximately parallel to the Honey Hill fault zone where the folds are overturned to the northwest. The eastern parts swing southeast across southwestern Rhode Island and is overturned to the northeast. Both the northwest- and northeast-dipping overturned folds and associated thrust faults merge and appear to have formed at the same time.

The West Rhode Island Fold Belt ends to the north just inside Massachusetts, where the Nashoba thrust belt changes direction to the northeast (Fig. 1). Farther north, the batholith is strongly foliated and sheared adjacent to the thrust belt rather than folded (sheared batholithic complex along the Bloody Bluff fault is seen on trip C4 by Barosh). The apparent stratigraphic units formed by the pendants within the batholitic complex trend west-southwest and are cut off obliquely at the Bloody Bluff fault zone that marks the southeast edge of the thrust belt.

Several northeast-trending aeromagnetic and gravity lineaments cross the Rhode Island Dome. The distribution pattern and attitude of the Blackstone Series where crossed by these geophysical lineaments indicates that they are fault zones with a few kilometers of right-lateral offset each. One of these, the Watch Hill lineament, extends at least through southwestern Rhode Island as a fault zone (Hermes, Barosh and Smith, 1981), and projects into the Narragansett Bay where the bay changes shape (Fig. 2). It and several geophysically defined faults in the bay area may form a major northeast-trending zone of an echelon faults that continue northeastward across the bay and through Fall River, Massachusetts. Another northeast-trending aeromagnetic and gravity lineament crosses northern Rhode Island. (Barosh, Pease, and others, 1977). This offset on this lineament, indicating a fault, is supported by ground geophysical studies (Schwab and Frohlich, 1976), and by the approximate alignment of the lineament with a fault in Narragansett Basin (Fig. 2).

The major geophysical features in the poorly exposed areas southeast of the Watch Hill lineament trend north-north-east (R.K. Frohlich, oral comm.) as they also do in the Narragansett Bay and off-shore. This north-east direction appears to represent the major structural trend in this area (Collins and McMaster, 1978; McMaster and others, 1980). The structural grain north of the lineament appears to be north to north-northwest, as expressed by the trend of contracts and a few known faults.

The northern on-shore part of the Southeast New England Platform is cut by numerous faults that were formed and commonly re-activated over a long period of time. Most of the faults trend between northeast and east and relatively younger northerly trending faults are common (Fig. 4).

Paleozoic plutons and both Paleozoic and Mesozoic dikes cut the platform (Mafic dikes north of Boston are visited on trip A6 by Ross). Ordovician plutonic rock has invaded the Southeast New England Platform in

Fig. 3 Sketch map and cross section of the West Rhode Island Fold Belt (Barosh and Hermes, 1981)

places and forms most of the area around Cape Ann, north of the Boston Basin. The Cape Ann Granite, which consists of several phases of quartz monzonite, is approximately coevel with the adjacent highly-variable Salem Gabbro-Diorite (Dennen, 1976, 1981). These rocks are cut off to the west by the Bloody Bluff fault system, but display none of the intense foliation and shearing near the fault that the older batholithic rock does. The Ordovician intrusive rock south of Boston is well represented by the Quincy Granite, which is non-foliated and fresh looking.

Some intrusive rock within the batholithic complex in Rhode Island has recently been found to be Devonian in age; the Cowesett Granite and some within the Scituate Granite Gneiss to the west (Hermes, Gromet and Zartman, 1981) (Fig. 2). However, part of the batholithic complex in southwestern Rhode Island previously dated as Devonian (Moore, 1959) is found to consist of a mixture of late Precambrian and Permian rock.

The southern shore of Rhode Island is underlain by the Narragansett Pier Granite and its aplitic phase the Westerly Granite of Permian age (Quinn, 1971) (Fig. 2). This non-foliated granite and its associated pegmatite are found cutting the Precambrian rock for a considerable distance to the north (Hermes, Barosh and Smith, 1981).

In addition, a large basic intrusion of unknown age with a northward elongation, is interpreted from gravity and magnetic data to underlie the western side of Cape Cod (Barosh, 1976), and magnetic highs off-shore to the south may indicate others (Barosh and others, 1977). The Silurian Preston Gabbro of southeast Connecticut (Zartman in Dixon, 1982) has similar magnetic expression and these may be the same age or they could possibly represent Mesozoic volcanoes.

Several basins lie within and at the edge of the platform. The Boston Basin is a structural basin filled with a wide variety of only very slightly metamorphosed coarse conglomerate, volcanic rock and argillite unconformably overlying the batholithic complex. The age of the rock ranges from latest Precambrian to Middle Cambrian (Kaye and Zartman, 1980) and possibly to Ordovician (Cambrian strata are seen on the north edge of the basin on trip C1 by Bailey). The rocks in the basin are broadly folded and highly faulted, as seen in the several tunnels across the basin (The rocks of the basin and its structure are examined on trip B2 by Kaye) (The southeast side of the basin and the adjacent batho-lithic complex are visited on trip C10 by Brenninkmeyer and Dillon). The nearly east-trending fault system controlling the Boston Basin continues westward across the batholitic rock complex and ends against the Bloody Bluff fault zone (Figs. 1, 4 and 5).

Fossiliferous Cambrian limestone and meta-shale also occurs outside of the basin as an isolated exposure within the northern Narragansett Basin (Shaler and others, 1899), and near its southern edge, where fossils were found recently by Trem Smith (Skehan and others, 1977) (Fig. 2). Volcanic rock similar to that in the Boston Basin is present on either side of the basin. The Lynn series to the north and that in the Blue Hills to the south (C.A. Kaye, oral comm.). The Lynn volcanic rock is formed of a wide variety of intermediate to acidic volcanic rock on the south side of Cape Ann. Well preserved ash-flow tuffs are prominent in

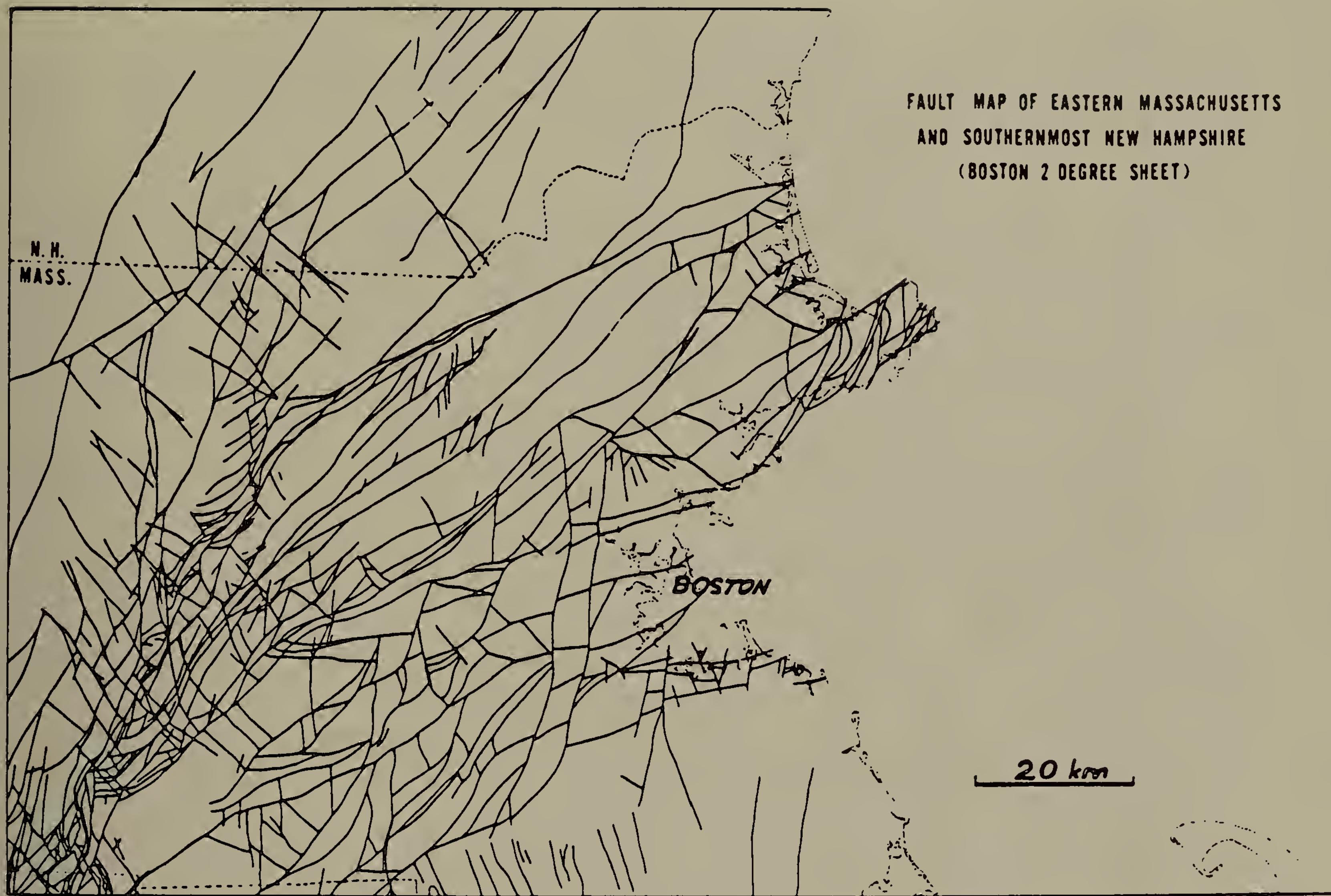

Fig. 4 Map of eastern Massachusetts and southern New Hampshire showing mapped faults (Barosh, Fahey and Pease, 1977)

Fig. 5 Sketch map showing structural relations of the Boston (B.B) and Norfolk Basins (N.B.) with the eastern edge of the Nashoba Thrust Belt.

the Blue Hills. These volcanic rocks appear related to Ordovician intrusive rocks that cut the batholith (Dennen, 1981, Warren, 1913, Sayer, 1974), but may possibly be a little older (Kaye and Zartman, 1980; Naylor, 1981).

The largely fault-bounded Norfolk and Narragansett Basins form prominent basins to the south of the Boston Basin (Figs. 1 and 2). They are filled mainly with nonmarine conglomerate, sandstone, shale, and some coal of Pennsylvanian age (Shaler and others, 1899; Lyons, 1978). Much of the conglomeratic rock at the borders are red beds. Rock of the Narragansett Basin is both folded and faulted. The northern part of the basin exhibits a few broad east-northeast-trending bands of rock interpreted as large open folds (Shaler and others, 1899) (Fig. 2), but recent gravity profiles suggest they may be due to fault blocks (Sherman, 1978). In contrast, small isoclinial to recumbent north-northeast-trending folds, some of which are overturned, locally occur in the southern part of the basin. These folds have been interpreted by some workers to represent multiple episodes of deformation (Burks and others, 1981). However, the folding appears to be highly irregular and local and the adjacent older rock does not exhibit these folds. There has been a great deal of movement along the coal seams in this area (John Rabin, oral comm., 1981) and folding above these thrust faults and the soft sediment deformation that is present can account for most features. The structure of the poorly exposed parts of the basin is not well known, but much of the west side is faulted. A northeast-trending fault apparently forms the boundary northeast of Fall River (Fig. 2), and the northern border is cut by numerous small north- to northwest-trending faults (Fig. 1). The basin ends to the south with an irregular border formed mainly by probable north-northeast- and northeast-trending faults near the mouth of Narragansett Bay (McMaster and others, 1980) (Fig. 6). The basin rocks are metamorphozed to upper amphibolite facies in the southwestern part of the basin and adjacent to the Permian Narragansett Pier Granite (Shaler and others, 1899; Quinn, 1971; Hermes, Barosh and Smith, 1981). The grade of metamorphism decreases to the north (Shaler and others, 1899; Quinn, 1971; and Hepburn and Rehmer, 1981). Isograds of this Permian metamorphic episode are truncated by the Narragansett Pier Granite, and the rocks were subjected to local retrograde metamorphism.

Graben basins of Late Triassic to Early Jurassic red clastic rock and basalt lie off-shore and buried beneath the thick glacial outwash sands and gravels of Cape Cod (Ballard and Uchupi, 1975) and have been reached by drill on Nantucket Island (Fogler and others, 1978) (Figs. 1 and 7). Other similar grabens lie farther off-shore to the north and northeast and across central Connecticut and Massachusetts to the west (at the western edge of Fig. 1). In addition, red conglomerate, sandstone and siltstone, that form a fault sliver within the Bloody Bluff fault system northwest of Boston have been found by C.A. Kaye (1983), to be of Late Triassic-Early Jurassic age (Fig. 7). Another similar nearby basin in the Bloody Bluff fault system contains Late Silurian or Early Devonian red congolomerate, sandstone, and siltstone mixed with a large assortment of volcanic rock, termed the Newbury Volcanic Complex (Shride, 1976). This unit of mixed marine and terrestrial rock is not known elsewhere in southeastern New England, but has similarities to rock of the same age in coastal southeastern Maine.

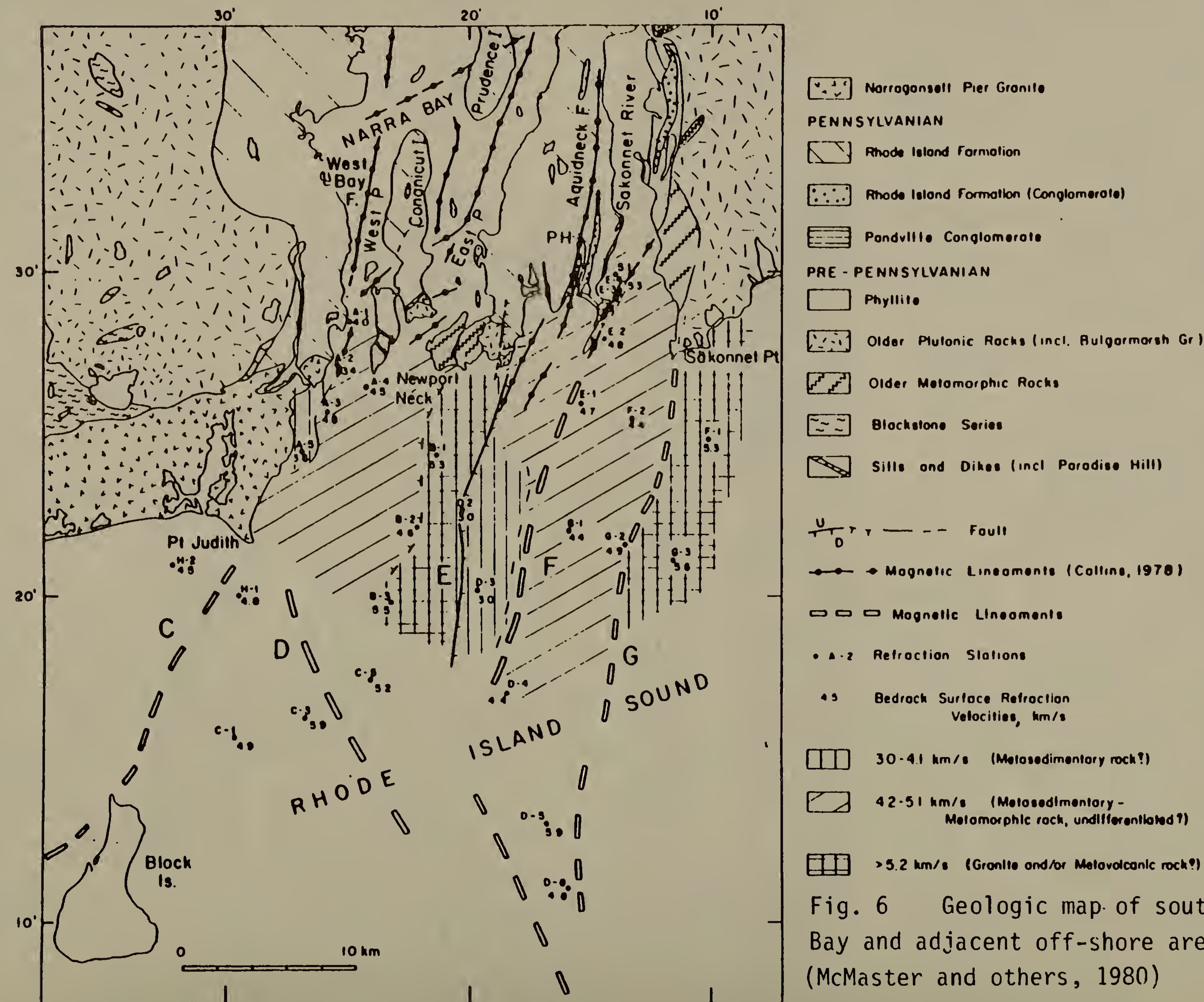

Fig. 6 Geologic map of southern Narraganset Bay and adjacent off-shore area (McMaster and others, 1980)

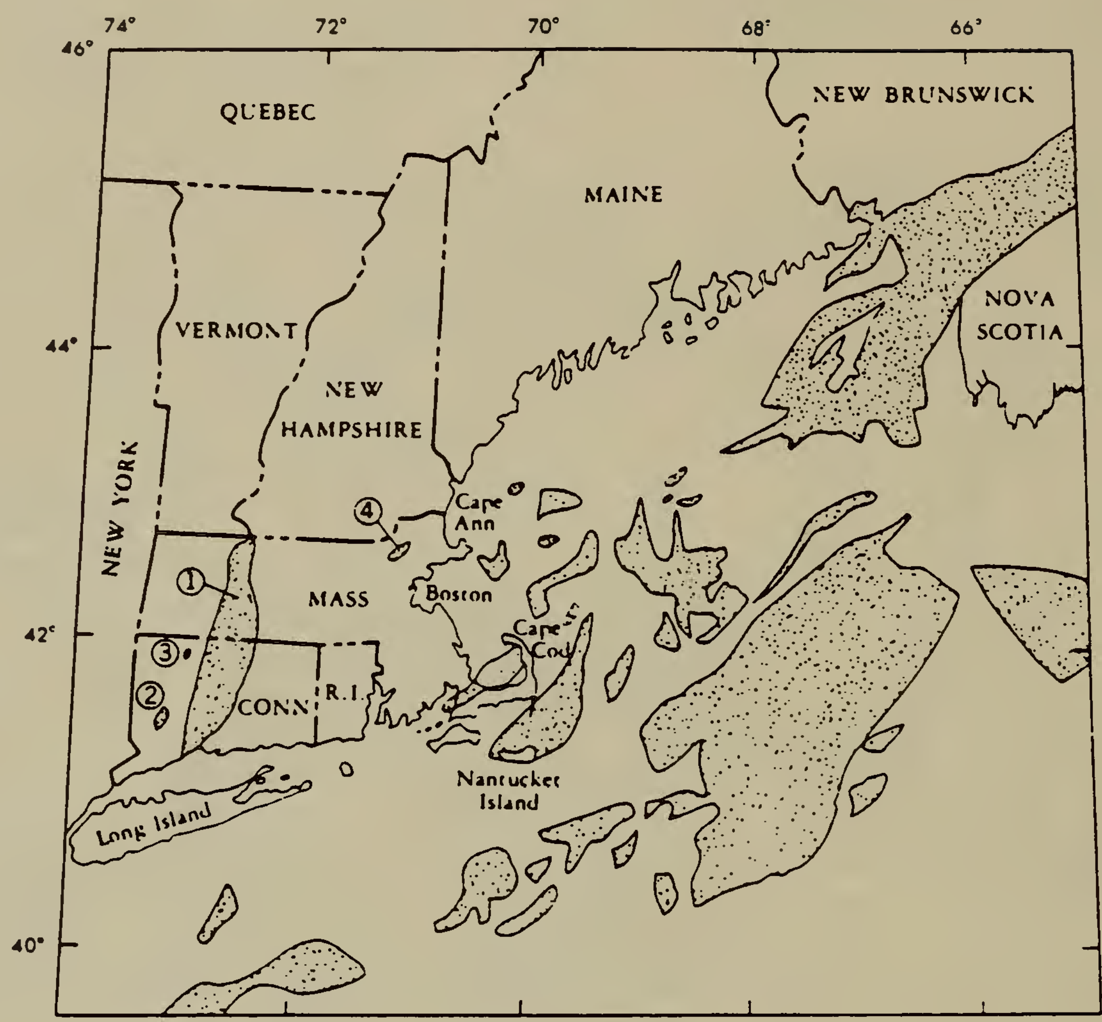

Fig. 7 Map of New England showing locations of basins containing the Late Triassic - Early Jurassic Newark Group (stippled). Off-shore basins from Ballard and Uchupi (1975). On-shore basins are the Hartford Graben (1), Pomperang Valley Graben (2), Canton Center Basin (3) and Middleton Basin (4) (from Kay, 1983)

A seaward thickening wedge of late Mesozoic and Tertiary deposits lies just off-shore and is part of the submerged northeast extension of the Atlantic Coastal Plain (Fig. 8). The Upper Cretaceous deposits offshore to the south form a northward-facing cuesta of sorts at the inner margin (O'Hara, oral comm., 1980). The consolidated Cretaceous clays and sands exposed at a few places on Block Island may be in place, but the Cretaceous and Tertiary deposits exposed at Martha's Vineyard to the east have been thrust up by glacial action (Kaye, 1964 a,b). Farther off-shore to the south and east a nearly continuous sequence from Middle Jurassic to upper Tertiary is present (Gibson and others, 1968; Valentine, 1981; Grow, 1981) (Fig. 8).

The small areas of Eocene sand and silt near the coast in the vicinity of Marshfield, south of Boston is the only outcrop of Tertiary now exposed on the mainland (Kaye, 1983), but others are known to lie just off-shore to the east (Weed, and others, 1974).

The Cretaceous is cut by a north to northwest-trending fault, the New Shoreham fault, off-shore south of Rhode Island (MacMaster, 1971). Some of the high-angle faults that cut the deposits at Marshfield and on Martha's Vineyard may be tectonic in origin and not related to glacial action (C.A. Kaye, pers. comm., 1983).

Sturbridge Geocline

The Sturbridge Geocline that lies in Connecticut, east-central Massachusetts, and southeastern New Hampshire to the west of the Southeast New England Platform and Nashoba Thrust Belt forms another major geologic province (Fig. 1). It is formed mainly of a very thick west-dipping and west-topping sequence of siltstone, graywacke, and shale. The sequence is known to be pre-Ordovician and is probably Precambrian in age (Barosh, 1981). These rocks have undergone high-grade metamorphism and are cut by numerous west-dipping thrust faults (Peper and others, 1975; Pease and Barosh, 1981).

This area has commonly been referred to as the Merrimack Geosyncline. However, it is not a geosyncline and has no structural connection with the structural trough containing the Devonian Littleton Formation in that was originally defined by Billings (1956), as the Merrimack geosyncline in north-central New Hampshire. Therefore, use of the term is confusing and should be abandoned. The Sturbridge Geocline consists of the generally northwest-dipping rock between the Nashoba Thrust Belt and the Bonemill Brook fault and other faults marking the east side of the intrusive cored uplifts of central Connecticut, Massachusetts, and southwest New Hampshire. It is named after Sturbridge, Massachusetts, which lies near the type sections of most of the meta-sedimentary units forming the province. The geocline is divided into western and eastern parts by the Nashoba Trough, a structural zone containing of relatively younger rock caught between steeply west-dipping thrust faults (Figs. 1 and 9).

The western part of the geocline consists of a very thick sequence of moderate to gentle northwest-dipping strata forming from base upwards the Oakdale formation, Paxton Group and Brimfield Group (Fig. 9). These units were defined near the turn of the century (Perry and Emerson,

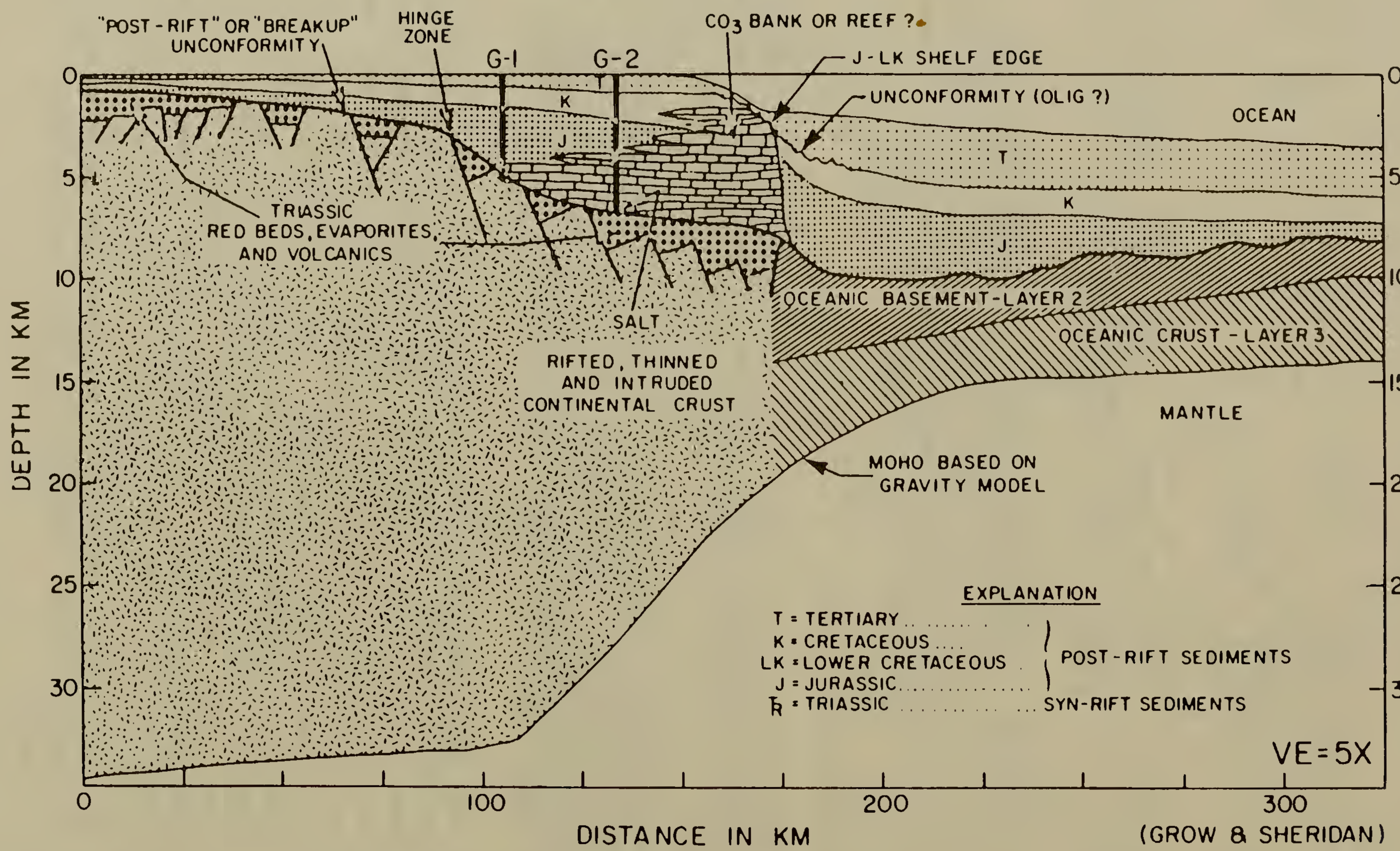

Fig. 8 Composite geologic section across southwest end of Georges Bank off-shore of southeastern New England (Grow, 1981)

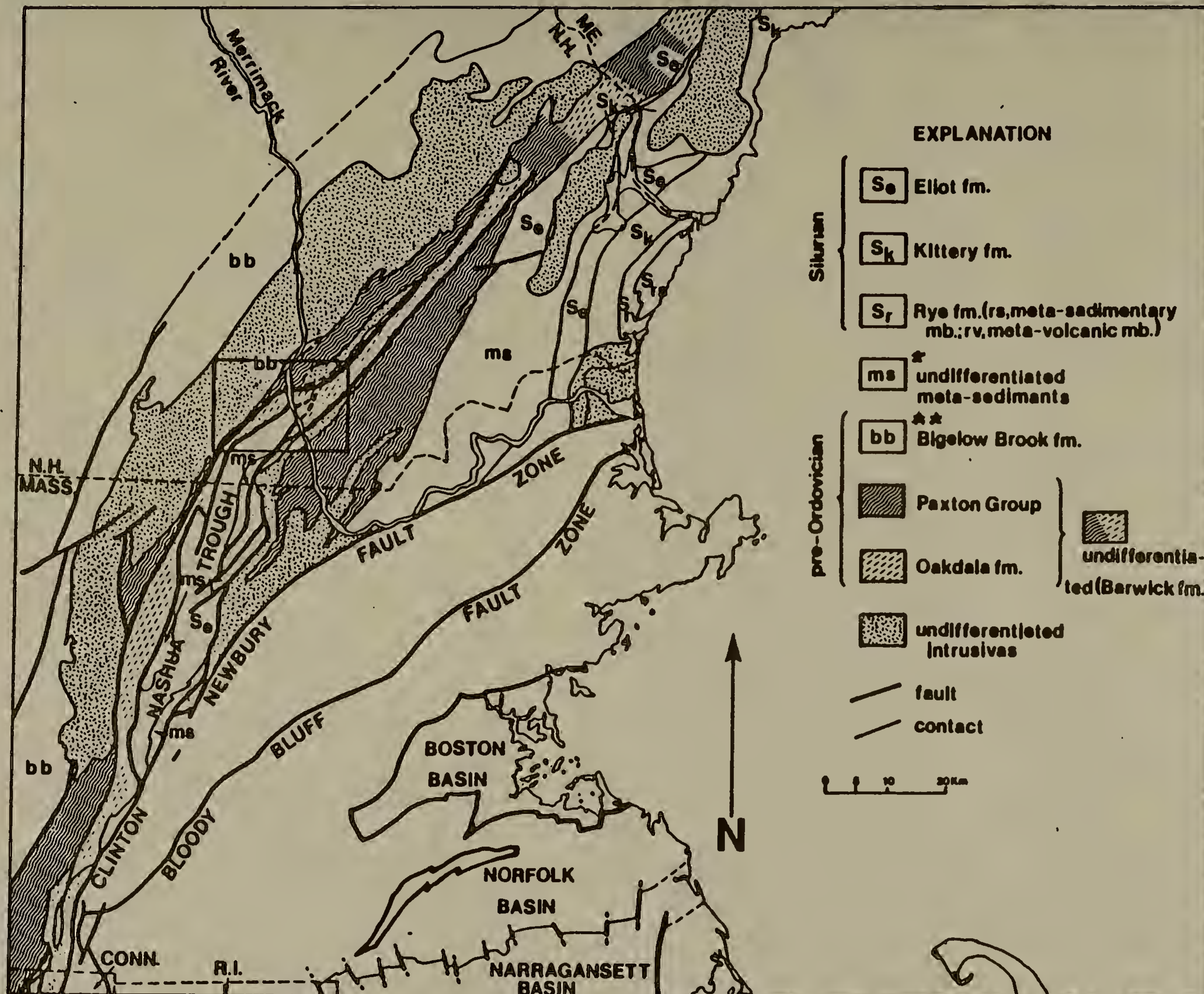

Fig. 9 Map of eastern Massachusetts and southeastern New Hampshire showing the bedrock geology of part of the Sturbridge Geocline (Smith and Barosh, 1982).

* MS is eastern block of Oakdale Formation

** Bigelow Brook Formation forms the base of the Brimfield Group

1903; Emerson, 1917) and have been painstakingly measured and redefined near the Massachusetts-Connecticut border in recent times (Pease, 1972; Peper and others, 1975, G.E. Moore, Jr. written commun. 1974; Barosh and others, 1977). They now have been followed southward to their terminous in east-central Connecticut and northeastward to southern Maine (Pease and Barosh, 1981; Barosh and Pease, 1981; Barosh and others, 1977) (Fig. 10) and provide the basis for understanding the structure of this geologic province. Other stratigraphic terms are still in use locally, such as the Berwick for the combined strata of the Oakdale and Paxton (The Berwick and adjacent rocks are seen on trip C5 by Eusden and others). The Shapleigh Group in Maine and much of the Littleton of southern New Hampshire are part of the Bigelow Brock and higher formations of the Brimfield Group (M.H. Pease, Jr., written commun. 1980) (Fig. 9).

The geocline is cut by numerous near bedding plane thrust faults, and a few high-angle ones in Connecticut. The thrust faults decrease in offset and die out to the north in Massachusetts, where a few very gentle folds are present, as shown in the Wachusetts water diversion tunnel across the region (Callaghan, 1931). The strata in Massachusetts and northward have been hypothesized to represent a series of very complex folded and refolded isoclinial folds (Thompson and others, 1968; Robinson and Hall, 1980), but several detailed stratigraphic studies and the continuous exposures in the tunnel indicate the limbs of the interpreted folds belong to different stratigraphic units (Peper and others, 1975; Pomeroy, 1974; Moore, 1976; Pease 1972; and Callaghan, 1931). No folds are observed in the area other than drag folds along thrust faults and the gentle folds seen in the tunnel.

Intrusive rock has invaded the early thrust faults at many locations, generally as elongated bodies. These mainly are quartz monzonite in composition and non-foliated to strongly foliated, depending largely upon whether or not movement on the intruded fault had stopped or continued. Most of the intrusions are pre-Late Silurian as they are older than the rock in the Nashoba Trough. The oldest, well-dated intrusive is Ordovician in age and cuts the Brimfield Group (Pease and Barosh, 1981). Another intrusion, the Massabesic Gneiss in southeast New Hampshire (Sriramadis, 1966), has been dated as late Precambrian (Besancon and others, 1977), but the sample area is highly contaminated by the invaded country rock and may be more representative of the age of the metasedimentary rocks than the intrusive. This migmatitic border zone forms the "gneissic" part of the Massabesic that otherwise appears less foliated than nearby intrusive rock considered Paleozoic in age. The Massabesic cuts both the Brimfield and Paxton Groups; the age of the meta-sedimentary sequence is therefore pre-Ordovician and probably Precambrian (Barosh and others, 1977, Barosh and Pease, 1981) (Some of these rocks adjacent to the Nashua Trough and important northwest-trending structures are seen on Trip C6 by Whitaker).

The Paxton Group and Oakdale Formation are repeated east of the Nashua trough by the major thrust faults bordering the trough (Figs. 1 and 9). The Oakdale Formation of this eastern block, referred to as "Eliot" by Sundeen (1971), also contains a thick muscovite schist lens equivalent to the thinner Gove member of the western block (Fig. 10). These

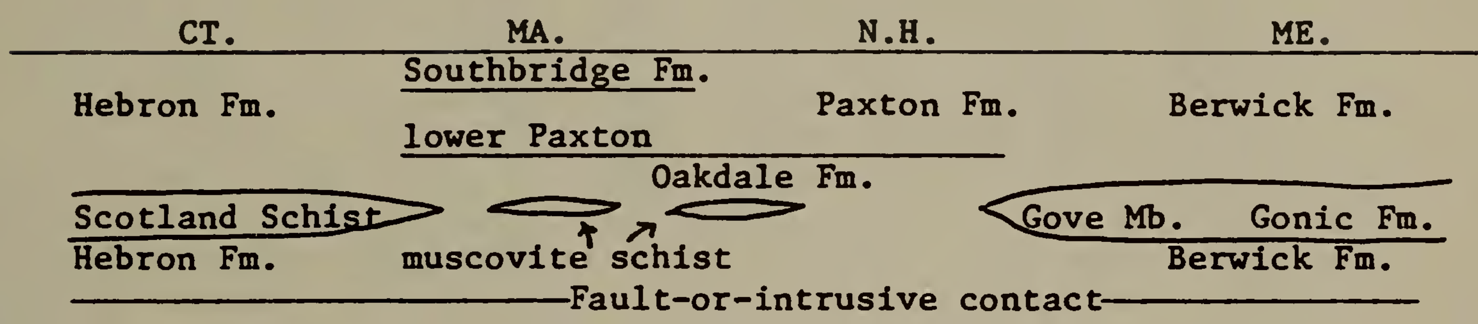

Fig. 10 Correlation chart for the Oakdale and Paxton Formations from Connecticut to southern Maine (Barosh and Pease, 1981). All overlain by the Brimfield Group.

formations are slightly finer grained than northwest of the trough and this suggests a source area to the northwest.

The Nashua Trough contains a sequence of rock consisting of thin-bedded quartzite, siltstone and shales, that are mildly metamorphized, described as units 1 to 4 by Peck (1976) (Smith and Barosh, 1981). These are all deposits formed on a submarine slope, distant from their source, by turbidity currents and, in some cases, slumps. Part of the sequence correlates with the Eliot Formation of southern Maine (Katz, 1918; Hussey, 1962). These are probably Late Silurian to Early Devonian in age. They were soft sediments and are locally isoclinaly folded near thrust faults. These folds are not known to extend into the adjacent older rock. The strata have not been seen to be intruded in this area. The trough impinges to the south on the Nashoba thrust belt and is cut off by the Clinton-Newbury fault zone along its border. Slivers of the rock in the trough are found in the fault zone farther southwest near Worcester and at Webster, near the Connecticut border (Fig. 1), providing some of the evidence for right-lateral movement along the fault zone. These younger strata are missing from the Merrimack River northeastward to the Lamprey River in New Hampshire and only the older more metamorphased Oakdale Formation is present in the trough.

Detailed geologic mapping and geophysical surveys show the Nashua Trough is formed of steeply north-dipping high-angle thrust faults; no major folds are present, although drag folds are common (Fig. 11) (Smith and Barosh, 1981, 1982). The trough forms the southern end of the Lewiston fault zone, that crosses all of Maine (Fig. 12). The zone has been gradually revealed by gravity, LANDSAT and local geologic studies.

Some of the rock in the Nashua Trough, the Eliot Formation, also occurs farther east in another faulted trough (Barosh and others, 1977) (Fig. 9). The units east of the Eliot, the Kittery and Rye Formations, probably underlie it, but their exact stratigraphic position is uncertain (These and nearby rocks are seen on trip A4 by Hussey and others) (The Rye Formation is examined on trip B4 by Swanson and Carrigan).

A volcanic chain developed in the Jurassic and Early Cretaceous along eastern New Hampshire and possibly off-shore to the edge of Cape Ann (One of the volcanoes, Mount Pawtuckaway, is examined on trip B7 by Eby).

High-angle northwest- and north-trending faults cut the northeast-trending thrust faults and are the youngest in the area (Figs. 4 and 11). Some are Late Cretaceous or younger as they cut upper Mesozoic volcanic complexes (Freedman, 1950).

Nashoba Thrust Belt

The Nashoba Thrust Belt is a major structural zone of closely spaced, steeply west-dipping thrust faults that separates the Southeast New England Platform from the Sturbridge Geocline (Figs 1 and 13.). The rock and structure of both these provinces terminate against it. No stratigraphic correlation has been possible between the pre-Ordovician

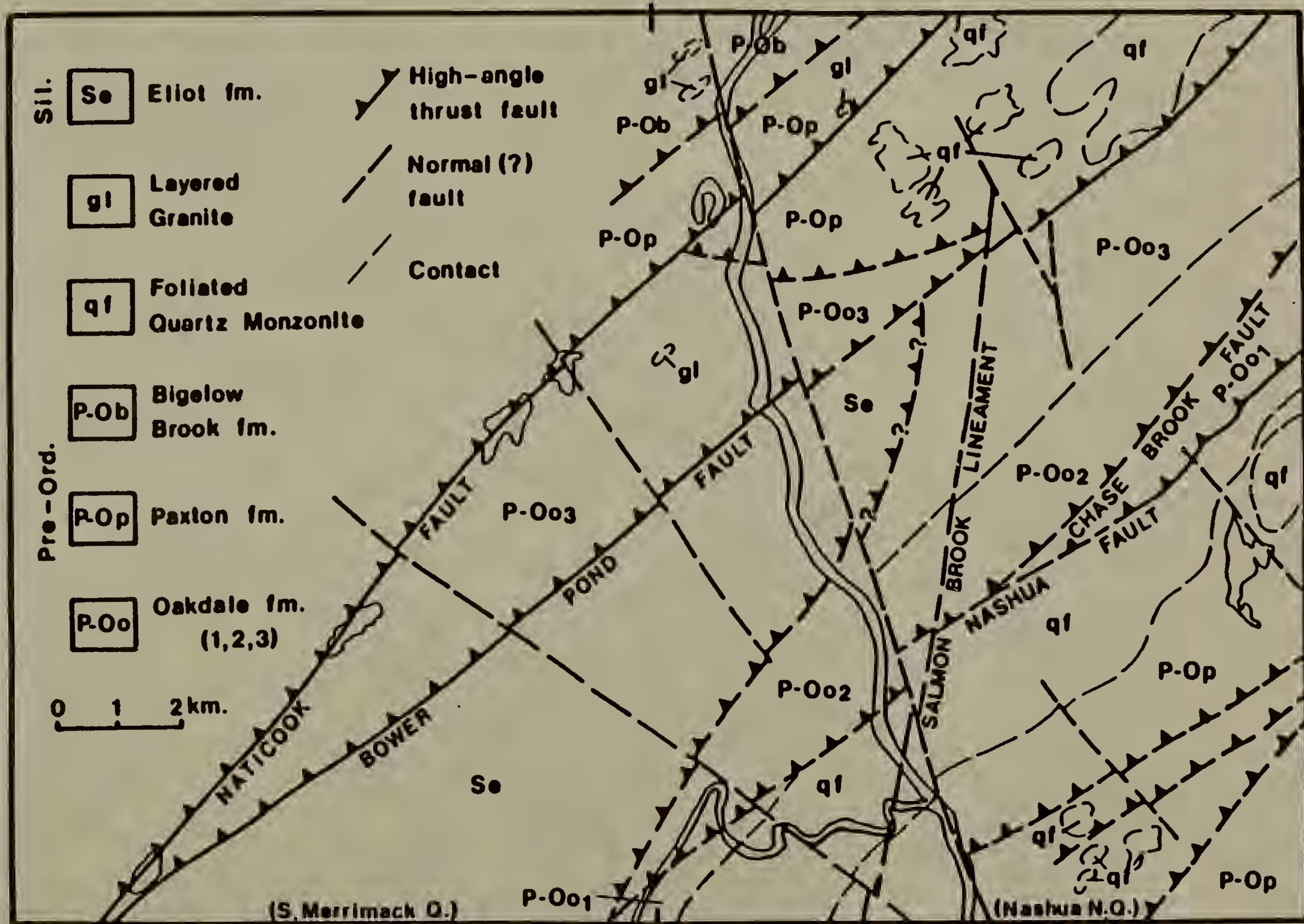

Fig. 11 Geologic sketch map of the Nashua North and South Merrimack 7.5-minute quadrangles, New Hampshire (Smith and Barosh, 1982). Location shown in figure 9.

rocks of these fault-bounded provinces and the rock in them probably formed at a considerable distance from one another.

The Nashoba Thrust Belt is wide in northeastern Massachusetts and contains nearly 18 kilometers of high metamorphic grade volcaniclastic rock measured northwest of Boston (Bell and Alvord, 1976, Alvord and others, 1976; Alvord, Pease and Fahey, 1976) (Fig. 13). This belt narrows drastically to the southwest, due to omission by thrust faults, and only a few hundred meters are present near Webster (Barosh, 1982) (Fig. 1); it widens again in eastern Connecticut. Many of the thrust zones in the belt are invaded by early to middle Paleozoic granitic rock, largely anatectic in origin. The units in the belt form a volcaniclastic sequence that consist mainly of thin, well-bedded amphibolite at the exposed base, a mixture of light-gray gneiss, schist, amphibolite, and marble in the middle and sillimanite-muscovite schist at the top. The units, known respectively as the Marlboro Formation (Quinebaug in CT.), Nashoba Formation of Hanson (Tatnic Hill in CT.), and the Tadmuck Brook Schist (not known in Connecticut), are described by Bell and Alvord (1976) (The Marlboro Formation is seen on trip C2 by DiNitto and others). These strata are invaded by a series of Ordovician (?) and Silurian (?) intrusive rocks and are probably Precambrian in age (The intrusive rocks are examined on trip A5 by Hill and others).

The principal movement along the northeast-trending thrust faults is west over east with a right-lateral component. These faults are cut by several small north- and northwest-trending high-angle faults (Fig. 4 and Fig. 4, Barosh, Trip C4)). Fault slivers of younger rock also occur along the borders of the Nashoba thrust belt in Massachusetts and attest to repeated movements along the thrust belt. Late Silurian (?) -Early Devonian (?) distal turbidites and Middle Pennsylvanian argillites occur in the Clinton-Newbury zone on its west side (Barosh, 1977) and Late Silurian-Early Devonian volcanic and sedimentary rock (Shride, 1976) and Late Triassic-Early Jurassic sedimentary rock in the Bloody Bluff fault system on its east side (the Bloody Bluff fault system and associated rock is examined on trip C4 by Barosh).

Tectonic Setting and Geologic History

The principal tectonic development of the region is believed to have taken place along the colliding border of two continental plates. The general concept of plate movements in New England was first mentioned by Wilson (1966) and elaborated on by Bird and Dewey (1970). Subsequent work has provided information on the structures involved and the timing of events. The movements appear to have occurred much earlier than first conceived. The Nashoba Thrust Belt apparently represents the west-dipping subduction zone between the Southeast New England Platform, a fragment of a former Paleo-African plate on the east, and the Sturbridge Geocline, a foreland basin of the North American plate (Barosh, 1979). These two plates probably moved towards one another obliquely along a path oriented east-northeast - west-southwest, as suggested by strain analysis of fault patterns (Barosh, 1976) and Paleozoic dikes and joints (Dennen, 1981). If the thick probable Precambrian strata of all three provinces in the region are broadly contemporaneous, then during the latter part of the Precambrian the following may have occurred. Muddy sand and sulfitic mud were carried southeastward and deposited in

Fig. 12 Map of the northeastern U.S. and adjacent Canada showing recent earthquakes and the Lewiston gravity lineament and other selected interpreted structures found or greatly extended on the basis of magnetic and gravity data (Barosh, 1982). 1. Northern Fall Line; 2. Connecticut River lineament; 3. Watch Hill lineament; 4. North Scituate-Blackstone lineament zone; 5. Higganum dike system; 6. Clinton-Newbury fault zone; 7. Winnipesaukee-Winooski lineament zone; 8. Penobscot lineament; 9. Norumbega fault zone; and 10. Upsalquitch lineament zone. Some geographical locations are: S, Sebago Lake; L, Lewiston; M, Medford; SL, South Lincoln; P, Passamaquoddy Bay; C, Chiputneticook Lakes; MB, Miramichi Bay.

a marine foreland basin now forming the Sturbridge Geocline. These sediments may have been derived from an inner volcanic arc, perhaps represented by some of the intrusive-cored domes of central Connecticut, Massachusetts, and western New Hampshire; an arc that stood off-shore east of the North American craton as reshaped by the 1,000 m.y. Grenville orgogony. Farther to the east andesitic and basaltic debris, along with some lime and mud accumulated around an outer volcanic arc. Across the sea to the east lay the edge of the Paleo-African craton being buried by quartz-rich sand, succeeded by mud, basaltic debris, basaltic and rhyolitic tuff and tuffaceous mud.

As the Paleo-African plate moved against and under the North American one, the outer arc became torn apart and carried into the subduction zone leaving only remnants of volcaniclastic rock found in the Nashoba Thrust Belt.

The main movement appears to have been in the late Precambrian with successive and gradually diminishing pulses during the Paleozoic. The greatest deformation on the Southeast New England platform is in the Late Precambrian rocks along its western edge against the Nashoba Thrust Belt. Late Precambrian intrusive cored (600-620 m.y.) folds are present and the intrusive rock along the Lake Char and Bloody Bluff faults is intensely sheared, whereas the Ordovician Cape Ann Granite and Salem Gabbro-Diorite along the Bloody Bluff fault zone, although faulted, are not intensely sheared. The Nashoba Thrust Belt parallels the general configuration of the folds, suggesting that thrusting was contemporaneous with the folding, a syntectonic late Precambrian event. It is also possible that the configuration of the late Precambrian folds somehow acted as a surface upon which later thrusting occurred, but this requires a number of coincidences. The major regional metamorphism occurred at this time on the platform and probably to the west as well.

Closely following the emplacement of the late Precambrian plutons and the accompanying tectonic activity, there was considerable uplift and erosion, perhaps due to isostatic rebound following the thickening of the crust in the subduction zone. The Boston basin was initiated in the latest Precambrian possibly due to extension caused by continued movement along the adjacent thrust belt (Fig. 5). The basin filled with a sequence of latest Precambrian near-shore volcanic rock and conglomerate interbedded with argillite, that grades upward to Middle Cambrian marine argillite (Kaye and Zartman, 1980). The rhyolitic and andesitic volcanic rock may reflect continuing activity along the western edge of the Paleo-African plate followed by a general transgressive sequence of off-shore muds and turbidites. Much of the rest of the platform was also covered by Cambrian sediments as indicated by scattered remnants. The coarse conglomerate and volcanic debris suggests high relief adjacent to the Boston Basin and probably basin and range structure existed then (C.A. Kaye, oral comm., 1982).

The Taconic orogeny affected the entire region and appears to have lasted over a longer period than in the western Appalachian Mountains, as suggested by the scattering of radio-metric age dates of associated plutons from mid Ordovician to mid Silurian. Thrusting took place in the Nashoba and Sturbridge terrains accompanied by formation of largely anatectic granite and pegmatite along the thrust faults. To the east,

Fig. 13 Geologic map of the central part of the Nashoba Thrust Belt in Massachusetts (Alvord and others, 1976). Intensive rock shown by stippling.

on the platform, the Cape Ann and Quincy Granite and Salem Gabbro-Diorite were intruded and accompanied by volcanism. The ash-flow tuffs in the Blue Hills probably came from a volcano represented by the Quincy Granite, although it is possible that they are slightly older (Naylor, 1981). The metamorphism accompanying the orogeny, although locally intense, is apparently less than previously occurred and the Boston Basin sediments were only very slightly affected.

Volcanic activity still affected the region in the Late Silurian or Early Devonian as shown by the variety of volcanic rock and red clastic rock mixed with marine sediments in the Newbury Volcanic Complex in north-eastern Massachusetts (Shride, 1976). Possibly, a volcanic chain connected them with the contemporaneous coastal volcanic sequence of eastern Maine. The two sites may have been much closer at that time and subsequently shifted farther apart by right-lateral movement along the Clinton-Newbury fault zone and others in the Nashoba Thrust Belt.

A deeper marine basin, into which sediments moved downslope southeastward in turbidity currents and occasional slumps, apparently lay farther northwest. The ridge from which similar southeastward moving sediments were derived passes through northwest Maine and northern New Hampshire (R. Moench, written comm., 1980) and probably continued through central Massachusetts and into Connecticut, perhaps farther away than the one that suppled the pre-Ordovician sediments.

Granitic rock intruded the region during the Acadian orogeny again over perhaps a wider age range than the traditional Middle Devonian date. Some cut the Sturbridge Geocline, especially in central New Hampshire and on the platform in Rhode Island. These generally alkalic rocks range in age from Early to Middle Devonian (Hermes, Gromet and Zartman, 1981). In Massachusetts, they include the Wenham Monzonite and the Rattlesnake Hill pluton (Lyons and Kruger, 1976). Southward in Rhode Island, alkalic rocks of the East Greenwich Group and parts of the Scituate Granite Gneiss yield Devonian ages (Hermes, Gromet and Zartman, 1981). These mid-Paleozoic plutons of the Southeastern New England platform, although generally contemporaneous with Acadian plutons in tectonic blocks to the west generally maintain a distinct petrologic character (Hermes, Gromet and Zartman, 1981). Moreover, the Spencer Hill volcanics of central Rhode Island have been interpreted by Quinn (1971) to be comagmatic with the nearby Devonian-aged Cowesett Granite. The Nashoba Trough does not appear to have been intruded, but the slight to moderate regional metamorphism that effected the rock in the trough (Peck, 1976) probably occurred at this time along with the development of local contact metamorphic aureoles.

Re-activation of the thrust faults in the Nashoba and Sturbridge provinces probably in the mid-Devonian caught up sedimentary rock in the Nashoba trough and other fault slivers and folded it locally during thrusting (Peck, 1976; Smith and Barosh, 1981). The thrusting continued to be west over east with a right-lateral component of movement.

The region may or may not have experienced the uplift and extensional faulting that led to the shedding of post-orogenic Late Devonian red clastic deposits in the coastal volcanic zone of eastern Maine (Schluger, 1973) and the Mississippian red beds in New Brunswick.

However, uplift, perhaps with associated extensional faulting, probably occurred on the Southeast New England Platform during the Pennsylvania to produce the non-marine conglomerate, sandstone, shale, and coal of the Narragansett and Norfolk Basins (Shaler and others, 1899). These deposits overlapped the Nashoba Thrust Belt as shown by the presence of a fault sliver of Pennsylvania rock on its west side in Worcester (Kemp, 1887; Grew, 1973). Fault scarps probably developed locally along the border of these basins during deposition, as indicated by the pebble to some very coarse conglomerates at the borders and the associated red sandstones and shales. These sediments are similar to those of the Late Devonian and Late Triassic-Early Jurassic grabens. The Norfolk Basin, in particular, may have formed similarly to that suggested for the Boston Basin by moment on the Bloody Bluff fault system (Fig. 5).

Metamorphism, intrusion and some deformation effected the southern edge of New England during the Permian Alleghenian orogeny. The sedimentary rock in the Narragansett Basin may have been deformed and metamorphosed mostly before the intrusion of the Narragansett Pier and Westerly Granites in the Permian (Burks and others, 1981; Hermes, Barosh and Smith, 1981), but these events are all part of the same tectonic episode. The highest grade of metamorphism roughly borders the southwestern margin of the basin adjacent to the granite and drops off to the north away from the granite. Illite crystallinity studies hint that two Alleghenian thermal events may have occurred (Hepburn and Rehmer, 1981). A number of recent studies indicate that the metamorphism may be more widespread than formerly realized (Zartman and others, 1970; Day and others, 1980; Dallmeyer, 1981). The metamorphism and intrusion are undoubtedly part of the same event with the high temperature just preceding the granite. The granite intruded in a relatively passive manner not disturbing the earlier structure of the rock (Hermes, Barosh and Smith, 1981).

Prior to the Late Triassic, the region was tilted to the north as shown by northward plunging structures and northward decrease in effects of different metamorphic events. Some of this tilt may have started early and contributed to the folding of the soft sediments at the south end of the Narragansett Basin.

Southern New England underwent uplift and extensional faulting during the Late Triassic and Early Jurassic as major rifting was initiated across the North Atlantic Basin. Deposition of continental clastic sediment and basalt occurred in local basins like that of the Hartford graben, that forms the Connecticut Valley (Hubert and others, 1978) and the Middleton basin (Kaye, 1983), but most basins now lie off-shore to the east (Ballard and Uchupi, 1975) (Fig. 7). Numerous diabase dikes were injected into the older rocks. Lamprophyric dikes also occur locally and may be of generally similar age (Ross, 1981). Continued extension into the Early Cretaceous probably helped create a volcanic chain that extended northerly, roughly parallel to the Hartford graben, along eastern New Hampshire. Normal movement appears to have occurred along many of the older thrust faults. Reactivation of the northeast-trending faults, that had a right-lateral component of movement during the Paleozoic, probably produced left-lateral movement (Ballard and Uchupi, 1975). However, there is no indication of any large-scale

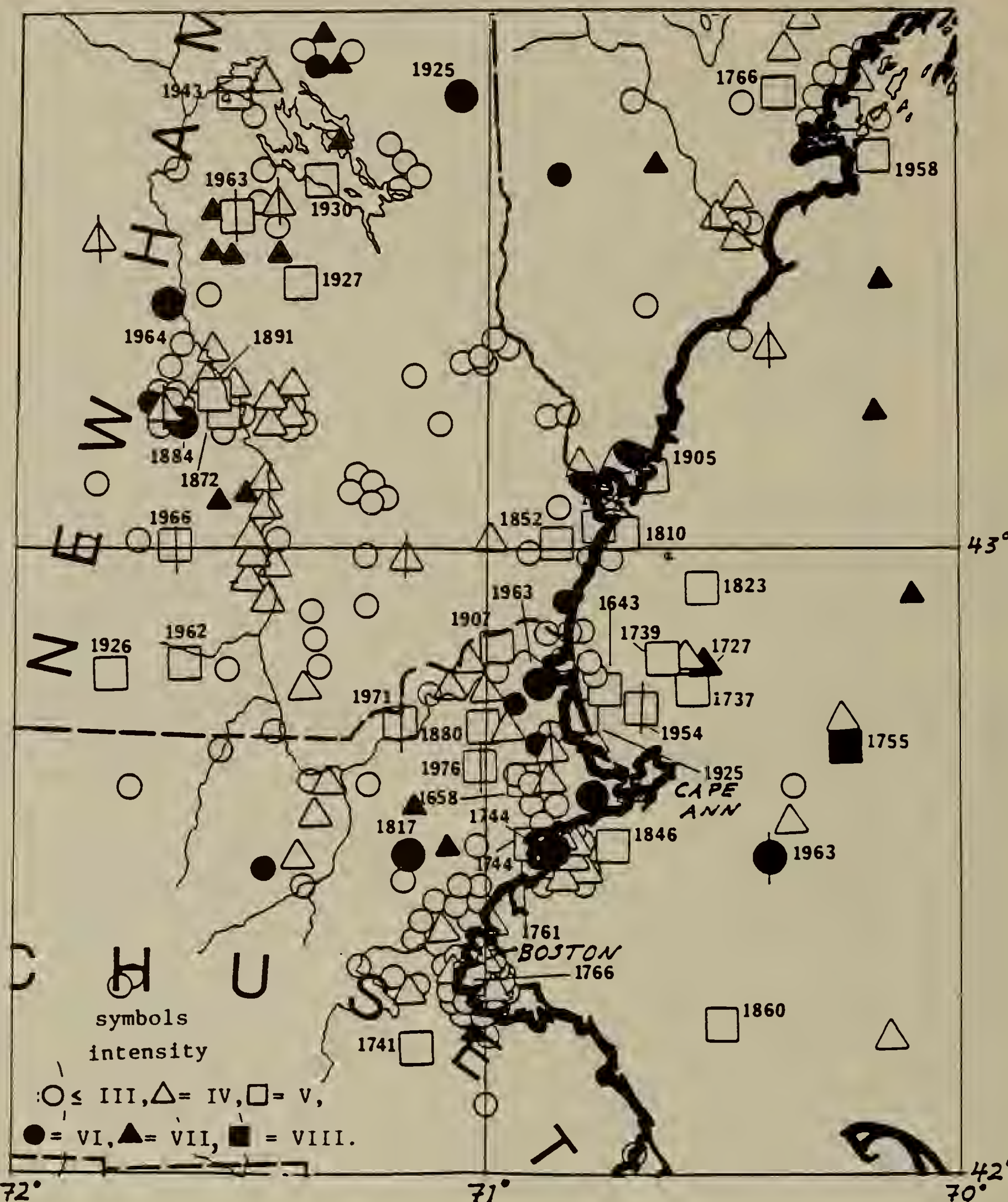

Fig. 14 Epicentral map of northeastern Massachusetts, southern New Hampshire and southern Maine through 1980 (from Nottis, 1983).

left-lateral movement such as that hypothesized by Kent and Opdyke (1978) from paleomagnetic data.

The southeastern edge of New England sagged to the south and east during the Middle Jurassic and Cretaceous as the North Atlantic Basin continued to open, and an apron of clastic sediments of Cretaceous age was deposited on it from the rising and eroding Appalachians to the west (Grow, 1981). Deposition continued into the Late Tertiary with an erosional hiatus during a low sea stand in the Oligocene (Weed and others, 1974; Valentine, 1981; and Kaye, 1983). Post-Cretaceous movements formed the north- to northwest-trending New Shoreham fault just west of Block Island (McMaster, 1971), and may have caused small movements, along many northwest-and some north-trending faults on shore.

Several periods of glaciation effected the region during the Pleistocene, but the only clear record is of the retreat of the late Wisconsin glacial cover. The entire region was covered by ice, including the near-shore area to the east and the region was depressed by the weight of the glacial ice. The rebound of the crust that began soon after the ice started its retreat 13,500 years ago has resulted in a regional tilt to the south of about 1m/km. This tilt and the post-glacial rise in sea level have caused the Late Pleistocene shore line to be deeply submerged off-shore to the south (USGS, 1976; O'Hara and Oldale, 1980), whereas it rises to the north above the present sea level just south of Boston.

Tectonic activity continues in the region as shown by the earthquakes around the Merrimack River Valley (Fig. 14), Cape Ann and Narragansett Bay. In the 1700's the Cape Ann area was much more active and two moderately large earthquakes occurred. These earthquakes appear to be due to local subsidence in the bays and river valley, that is related to continued opening of the North Atlantic basin (Barosh, 1981, in press). Short segments of northwest- and north-trending faults appear to be involved in this movement, particularly at fault intersections.

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QUATERNARY GEOLOGY AND GEOMORPHOLOGY

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The Quaternary geology of the coastal lowlands from Boston to Kennebunk, Maine is dominated by glaciation. Nearly all preglacial regolith was stripped off as glaciers scoured the landscape. Basins, such as the Boston Basin (Kaye, 1976), were excavated in regions underlain by less resistant rock, while preglacial valleys were widened and more deeply incised. Bedrock bosses were streamlined or rounded and an irregular blanket of till was deposited over the region. The drumlinoid topography characteristic of the Boston and Danvers areas (fig. 1) appears to be the product of at least two glaciations (Kaye, 1961, 1976, 1981; Oldale, 1964; Schafer and Hartshorn, 1965). In lowlands and valleys inland from the coast, ice retreat was accompanied by the deposition of glaciofluvial and glaciolaucustrine sediments (Mayewski and Birch, Trip C7; Koteff et al., Trip B9). Deglaciation of coastal regions was accompanied by a marine incursion which deposited a blanket of flocculated rock flour, or marine "clay". The thin deposits of marine clay, found as far south as Quincy, Massachusetts (Kaye, 1976), thicken and become more widespread northward into Maine. Figure 2 illustrates the areas of marine submergence, as determined from the occurrence of glaciomarine sediments. The thickness of glacial drift both onshore and offshore suggests that the average amount of glacial erosion may be 65 feet (approximately 20 meters) or more (Schafer and Hartshorn 1965).

Glacial History

Glacial events preceeding the early-Wisconsinan (Altonian) glaciation are poorly recorded in New England. Till deposits identified as pre-Wisconsinan (Kaye 1964a, 1964b; Oldale, 1982) are exceedingly rare and difficult to decipher. The scarcity of older drift probably attests to the highly erosive, cannibalistic power of the younger glaciers rather than to an absence of pre-Wisconsinan glaciation. If the weathered mantle formed through millions of years of exposure can be stripped off with barely a trace, the same may hold true for previous drift sheets.

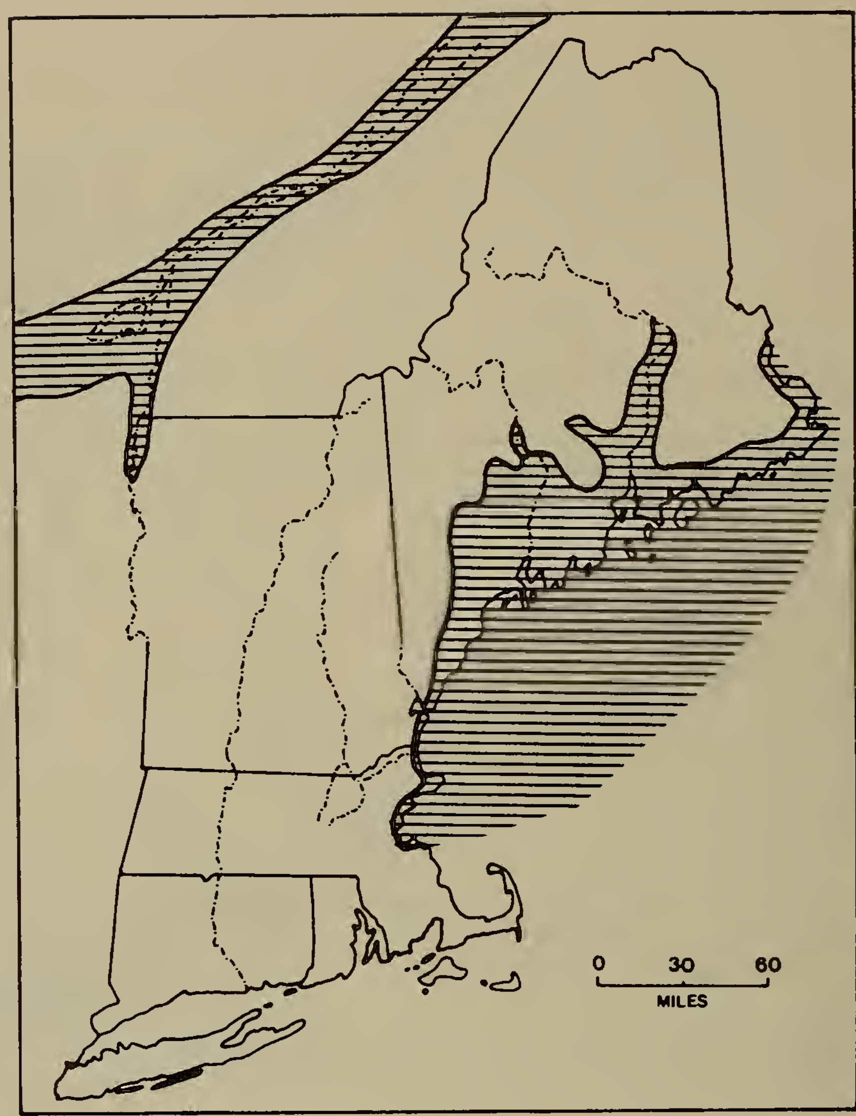
Stratigraphic evidence on Long Island and the islands south of Cape Cod indicates that the southeast margin of the early-Wisconsinan ice sheet terminated on the continental shelf (Kaye 1964a, 1964b; Oldale, 1982; Sirkin, 1982). By 75,000 years B.P. the ice had retreated from New England and the St. Lawrence Lowlands. This date corresponds to the St. Pierre beds (Stuiver, et al., 1978, McDonald and Shilts, 1971) which separate the early-Wisconsinan Becancour and middle-thru late- Wisconsinan Gentilly tills in the St. Lawrence Lowlands (fig. 3). The non-glacial, freshwater St. Pierre deposits signify that the region at that time was ice free.

The early-Wisconsinan ice sheet deposited a thick blanket of till which was subsequently eroded during the late-Wisconsinan (Woodfordian) substage. Many drumlins in Boston and Danvers contain a mantle of oxidized drift, 20 to 50 feet thick, and are believed to have been formed by differential erosion and remolding of older Altonian, or possibly pre-Wisconsinan, drift during the late-

Wisconsinan glaciation (Kaye, 1976, 1981; Oldale, 1964; Schafer and Hartshorn, 1965). Oldale (1964) suggests that the early-Wisconsinan drumlin till may be as thick as 200 feet in the Salem-Danvers area.

Figure 1. Map of the northeast coast of Massachusetts showing the drumlin fields of the Boston and Danvers areas. In part after Kaye (1976). Many of these drumlins are composed of late-Wisconsinan, or pre-Wisconinan, till which has been more deeply weathered than surrounding late-Wisconsinan drift. The gravel beaches of Wintrop and Nantasket are derived from erosion of adjacent drumlin headlands. Note that majority of islands in Boston Harbor, including Thompson Island (TI), are composed of drumlins.

Figure 2. Map of New England and Southern Quebec illustrating regions inundated by the late- and post-glacial marine submergence (Compiled from numerous sources). Note the marine submergence is much more extensive in Maine. The marine transgression into the St. Laurence Lowlands separated the Laurentide ice sheet, stranding ice southeastern Quebec and in Maine.



The southeastern ice margin of the Altonian ice sheet consisted of numerous lobes which thrust sediment and dumped debris along their termini, initiating the construction of Long Island and the islands south of Cape Cod (Sirkin, 1982; Oldale, 1982).

Retreat of the early-Wisconsinan glacier from southeastern New England was followed by a lengthy interstade during which Altonian drift deposits were deeply weathered. Middle-Wisconsinan pollen stratigraphy of sediments on Long Island outlines a fairly detailed account of climatic trends preceeding the late-Wisconsinan glacial advance (Sirkin and Stuckenrath, 1980, Sirkin, 1982). Many climatic events recorded in the pollen zones of Long Island and Block Island correlate remarkably well with glacial events to the northeast in southeastern Quebec and the St. Lawrence Lowlands (fig. 3). The dissipation of the early-Wisconsinan ice sheet was followed by a period of warming, recorded by temperate forest pollen dated older than 42,000 B.P. (Sirkin and Stuckenrath, 1980). The Nassauan Spruce Pollen Zone (Sirkin and Stuckenrath, 1980) records a subsequent cooling trend between 42,000 and 33,000 B.P. and although no evidence indicates a readvance into southeastern New England at this time a glacial advance into southeastern Quebec is recorded by the deposition of the middle-Wisconsinan Chaudiere Till (McDonald and Shilts, 1971). Following the Nassauan substage a warming trend occurred between 33,000 and 28,000 B.P. as evidenced by the Portwashington Oak Pollen Zone of Long Island (Sirkin and Stuckenrath, 1980). In southeastern Quebec this warming trend is reflected by a glacial retreat to the St. Lawrence Lowlands and deposition of the glaciolacustrine Gayhurst Formation in the Chaudiere and St. Francis river basins. These glacial-lake clays are overlain by the late-Wisconsinan Lennoxvile Till (McDonald and Shilts, 1971). Cooling between 28,000-21,750 B.P. preceeded the arrival of the late-Wisconsinan Laurentide ice sheet into southeastern New England and is recorded by the Farmdalian Spruce Pollen Zone (Sirkin and Stuckenrath, 1980).

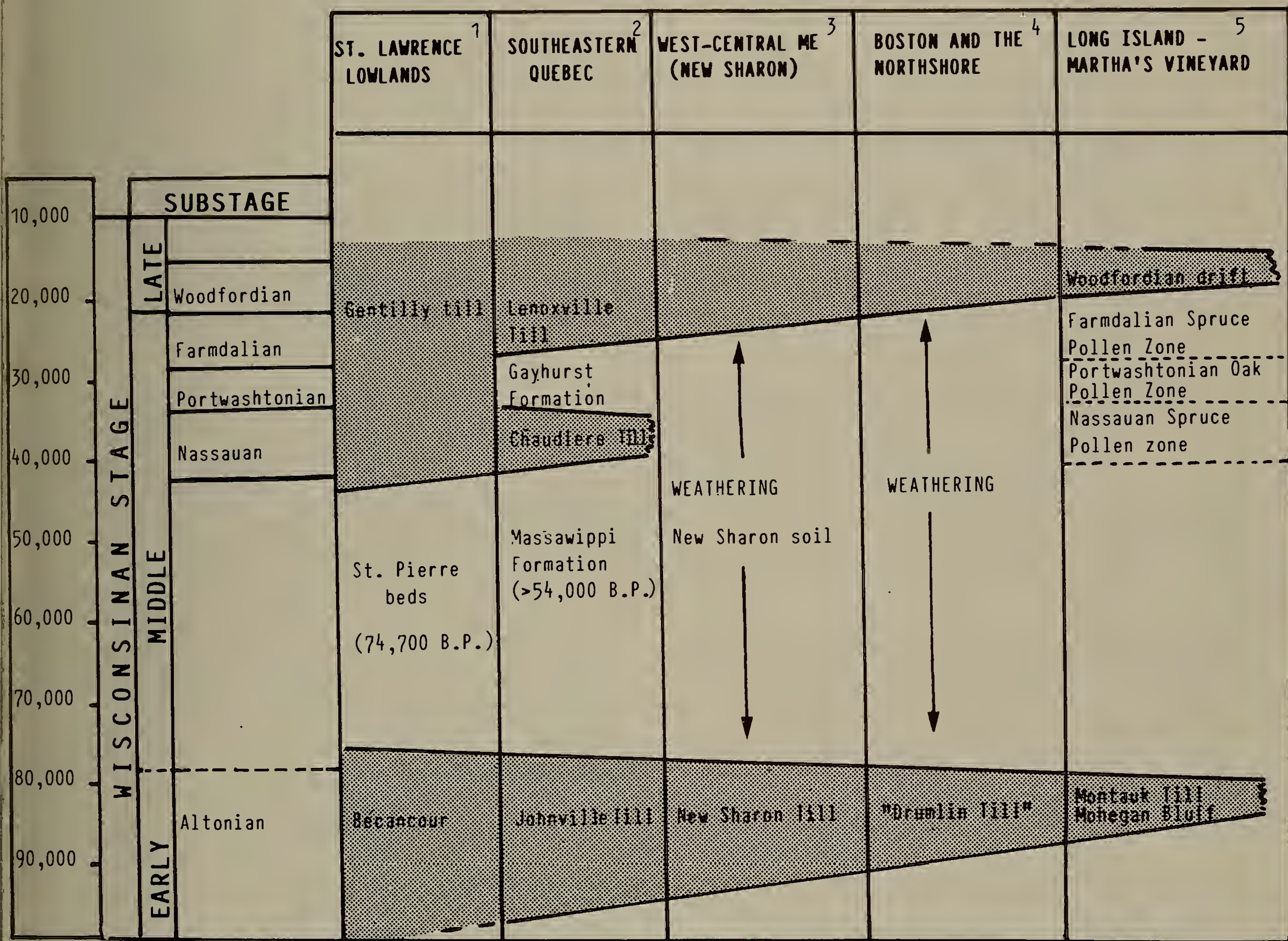


Figure 3. Time-distance diagram and stratigraphic chart correlating Wisconsinan substages, as defined by pollen stratigraphy on Long Island and Block Island (Sirken and Stukenrath, 1980, Sirkin, 1982), with glacial advances and retreats through southern Quebec and New England. Numbers refer to the following references: 1) McDonald and Shilts (1971); 2) Gadd (1971); 3) Caldwell (1959, 1960); 4) Odale (1964) and Kaye (1981); and 5) Sirkin and Stuckenrath (1980) and Sirkin (1982).

The Laurentide ice sheet which covered the region during the late Wisconsinan advanced southeastward through New England and reached its terminal position soon after 21,750 B.P. (Sirkin and Struckenrath, 1980). The ice margin was multilobate and occupied nearly the same position as the early-Wisconsinan glacier Long Island, Block Island, Martha's Vineyard and Nantucket are composed of the terminal moraines of both the early- and late-Wisconsinan glaciers (Sirkin, 1982; Oldale, 1982). In addition to Wisconsinan moraines, Kaye (1961b, 1961c) has found evidence of pre-Wisconsinan morainal deposits on Martha's Vineyard. Retreat of the Laurentide ice sheet from its terminal position on Martha's Vineyard occurred soon after 15,300 B.P. (Kaye, 1964b). The large recessional moraines and outwash plains of Cape Cod were deposited as oscillating ice lobes along southeastern ice margin followed a path of nonsynchronous retreat from the continental shelf into southeastern Massachusetts (Woodworth and Wigglesworth, 1934, Larson, 1982, Odale, 1982). Until the ice reached the Boston area its southeastern margin was predominantly land-based. Isostatic depression of coastal regions south of Quincy, Massachusetts was not sufficient to cause submergence beneath the prevailing sea level. Two dates from barnacle plates taken from marine clay in West Lynn show that the ice had retreated to the Northshore area by 14,000 B.P. (Kaye and Barghoorn, 1964, Kaye 1961, Stone and Peper, 1982).

In the Boston area Kaye (1982) has identified two late-Wisconsinan tills (Till III and IV). The older till (Till III) is the principal till of the late-Wisconsinan (Woodfordian) glaciation and is overlain by marine clay (Outwash III), commonly referred to as the "Boston blue clay". Locally, in the Back Bay and in Cambridge, a younger, enigmatic till, as thick as 35 feet (10m), overlies the clay and is thought to have been deposited by the Fresh Pond (Cambridge) readvance describe by Chute (1959). At least six dates, ranging in age from 12,200 to 12,700 B.P., on wood and peat collected from the underlying clay would place the readvance around 12,000 B.P. (Kaye, personal communication), at a time when most of New England is considered ice free. In explanation, Kaye suggests the possible existence of an ice mass in the uplands northwest of the Boston Basin. The structure, stratigraphic relationships, and wide-spread occurrence of the younger till in the Boston area, precludes the possibility that it may be a debris flow (Kaye, personal communication).

The sequence of events outlined by Oldale (1964) for the Salem area is quite different from that suggested by Kaye for the Boston area. Only one late-Wisconsinan till has been recognized and a lower relative sea level, concurrent with deglaciation of the coast, is inferred by the presence of low-level glaciofluvial sediments in coastal valleys. According to Oldale, further retreat was accompanied by a late-glacial submergence of low-lying coastal regions as the rise in sea level exceeded isostatic recovery. The submergence is recorded by marine clay found 50 feet or more above present sea level and high-level glaciofluvial deposits graded to the higher sea level. In contrast, glaciotectonic deformation of marine clays in Lynn attest to the proximity of the ice during submergence (Kaye, 1961; Stone and Peper, 1982). The deformed clays could be explained by a readvance, following a rise in sea level of over 50 feet, or by a re-interpretation of the low-level glaciofluvial sediments as discussed later in this introduction.

Soon after the onset of deglaciation, rapid calving of ice in the Gulf of Maine brought the glacial margin close to the Maine coast (Fastook and Hughes, 1982; Thompson, 1982). In southern Maine the ice had withdrawn from its terminal postion on the continental shelf to the coast by 13,500 B.P. (Stuiver and Borns, 1975), or possibly as early as 14,000 B.P. (Smith, Trip A7), where it remained until approximately 13,200 B.P..

Stratigraphic evidence recorded by numerous workers in Maine indicates that ice withdrawal from the coast was contemperaneous with marine submergence. However, Bloom (1960 and 1963) reported well sorted stratified drift, described in engineering testhole logs, 62-68 feet beneath present sea level and overlain by marine clay. He assumed these sediments to be subaerial-fluvial deposits based on their well-sorted character and postulated an emergence of approximately 70 feet prior to the marine transgression. Understanding of glaciomarine lithofacies, as outlined by Smith (Trip A7) can be invaluable to the interpretation of such sediments. The fluvial sediments described by Bloom are better interpreted as proximal subaqueous outwash formed by the retreat of a marine-based glacier. Such a re-evaluation of the low-level glaciofluvial deposits along the Northshore in Massachusetts may also be in order.

While ice still occupied coastal Maine, between 14,000 and 13,200 B.P., deglaciation of Massachusetts and southern New Hampshire was nearly complete (Caldwell et al., 1978; Koteff et al., Trip B9). Stratigraphic evidence throughout New England indicates that deglaciation was accomplished by active-ice, stagnation-zone retreat and not by regional stagnation (Koteff, 1974; Koteff and Pessl, 1981). Active-ice retreat produced sucessive sequences, termed morphosequences, of ice-contact and proglacial sediments which can be used to determine consecutive ice-marginal positions (Koteff, 1974; Koteff and Pessl 1981, and Koteff et al., Trip B9). Koteff et al (Trip B9) discusses morphosequences formed in a glaciolacustrine environment while tracing the retreat of ice up the Merrimac Valley. Mayewski et al. (Trip C7) illustrate evidence for active- ice retreat from coastal New Hampshire. Morphosequences and lithofacies formed in a glaciomarine environment are outlined by Smith (Trip A7).

As the marine transgression accompanied withdrawal of ice from coastal Maine and New Hampshire a variety of glaciomarine sediment were deposited, including a thick blanket of marine clay known in Maine as the Presumscott Formation (Bloom, 1960). Glaciomarine deltas graded to the higher relative sea level form the broad elevated plains commonly seen in the coastal lowlands (Trips A7 and B7). The blueberry industry of Maine owes its existence to the abundant, well drained, stratified glaciomarine sediments formed during this stage of deglaciation.

By 12,400 B.P. the ice had retreated inland from limit of marine submergence (Stuiver and Borns, 1975; Smith, Trip A7). Withdrawal of ice from coastal Maine was contemporaneous with the formation of a calving bay in the Gulf of St. Lawrence (McDonald, 1968; Borns 1963, 1966, 1967; Stuiver and Borns 1975). A later transgression of the Champlain Sea into the St. Lawrence Lowlands cut into the Laurentide ice sheet and stranded ice in southeastern Quebec and Maine (Shilts, 1976; Hanson 1977). Rapid dissipation of the ice soon followed. By 11,500 B.P isostatic rebound and consequent marine regression along the coast was complete (Stuiver et. al, 1971; Smith, Trip A7). The retreat of marine waters from submerged portions of the coast is recorded by numerous sequential beach ridges and wave-cut escarpments (Smith, Trip A7).

COASTAL GEOLOGY

Bedrock geology, glacial history, sediment supply, hydrographic regime, and post-glacial sea-level fluctuations all play an important role in the construction and modification of the present shoreline. The ragged outline of the coast from Boston northward, seen in figure 1, is largely the result of differential erosion and scouring of bedrock during glaciation and subsequent submergence. Major headlands composed of resistant rock surround embayments underlain by weaker rock. The headlands of Cape Ann and Salem are underlain by the more resistant Cape Ann Granite and Salem Gabbro-Diorite. Directly to the south, lies the Boston embayment underlain by the generally weaker, slightly metamorphosed, sedimentary rocks of the Boston Basin. Highly faulted, metamorphosed Precambrian and Paleozoic rocks form the embayment north of Cape Ann.

The preglacial shoreline, exposed to weathering and coastal processes for millions of years, was undoubtably smoother than the present shore and fringed with well developed beaches. These more mature beach sediments were carried offshore onto the continental shelf while glaciers scoured the underlying bedrock and deposited irregular sheets of till, sand and gravel.

Figure 4. Nasa Landsat image of of the coast from Boston Massachusetts to York Harbor, Maine.

Drumlins, minor glaciofluvial deposits, and bedrock supply most of the beach sediments from Scituate to Cape Ann. Drumlins dominate the landscape in the Boston area and along the north and south shores. In Boston Harbor, eroding drumlins form small islands and shoals. A till drumlin and gravel outwash plain are combined to form Thompson Island, one of the larger islands in the harbor (Caldwell, Trip C8). Erosion and redeposition of these glacial deposits by waves has produced a variety of gravel spits on the island (Rosen, Trip A1). Johnson (1910, 1919, 1925) illustrated how erosion and redistribution of drumlin-derived sediment into gravel tombolos and spits lead to the construction of the Winthrop and Nantasket barrrier beaches (Brenninkmeyer and Dillon, Trip C10; FitzGerald, Trip A1).

The large sandy barrier islands of Plum Island and Castle Neck are nourished by the Merrimack River, as it erodes through the thick blanket of glaciofluvial sediments left in it's valley. Whether these barrier beaches were created from spits, high dune ridges, or offshore bars which migrated shoreward during the Holocene transgression has been the subject of much controversy. For further discussion see McIntire and Morgan (1963) or Jones and Cameron (1976).

Hydrographic regime governs the patterns of sediment transport and deposition that contol the shape and dimensions of the barrier beaches, their intervening inlets, and tidal deltas. The coast from Boston to Kennebunk is predominantly a mixed energy coast (Hayes, 1979) whereby tides and waves maintain nearly equally important roles in sediment transport and deposition (see FitzGerald, Trip A1). Barrier beaches are short, backed by marsh and tidal creek systems, and are associated with well developed flood and ebb tidal deltas.

The rise in relative sea level has resulted in the landward migration of barrier beaches for the past several thousand years. The shoreward migration of barrier beaches is a cannibalistic processes. Overwash, tidal, and eolian processes transport sediment from the foreshore to the back-barrier environment, thereby enabling landward migration and preservation with rising sea level. According to Kaye and Barghoorn (1964), sea level rose from a minimum of -70 feet, at 10,000 B.P., to approximately its present position by 2,000 B.P.. Since this time sea level has been fluctuating to within 1.5 feet of present sea level. Using barnacles as an indicator of recent sea level fluctuations, Kaye (1964) observed that between 1922 and 1931, following a fall in sea level of .17 feet since 1856, sea level was -.07 feet. By 1961 sea level has risen .72 feet and is continuing to rise. The rapid rise in sea level in recent years has lead to accelerated erosion and destruction in coastal communities. Many barrier beaches, such as those in Revere, Lynn, and Swampscott, have been over developed. The construction of seawalls and other coastal structures, which decrease foreshore erosion and back-barrier sediment transport. may ultimately lead the destruction of these barrier beaches if sea level continues to rise. FitzGerald (Trip A1) discusses the effectiveness and consequences of various coastal structures along the Winthrop shore.

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